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FINAL REPORT

The Aggregate Representation of Terrestrial Land Covers Within Global Climate Models (GCM)

(NASA Contract No. NAGW-3368)

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1. Overview of Activities and Progress

This project had four initial objectives:

1. to create a realistic coupled surface-atmosphere model to investigate the aggregate description of heterogeneous surfaces;
2. to develop a simple heuristic model of surface-atmosphere interactions;
3. using the above models, to test aggregation rules for a variety of realistic cover and meteorological conditions; and
4. to reconcile BATS land covers with those that can be recognized from space;

Our progress in meeting these objectives can be summarized as follows.

Objective 1: The first objective was achieved in the first year of the project by coupling the Biosphere-Atmosphere Transfer Scheme (BATS) with a proven two-dimensional model of the atmospheric boundary layer. The resulting model, BATS-ABL, is described in detail in a Masters thesis (Arain, 1994) and reported in a paper in the Journal of Hydrology (Arain *et al.*, 1996).

Objective 2: The potential value of the heuristic model was re-evaluated early in the project and a decision was made to focus subsequent research around modeling studies with the BATS-ABL model (see first annual report). Given the subsequent success in applying this last model this was clearly a wise decision. The value of using such coupled surface-atmosphere models in this research area was further confirmed by the success of the Tucson Aggregation Workshop (see below).

Objective 3: There was excellent progress in using the BATS-ABL model to test aggregation rules for a variety of realistic covers. The foci of attention have been the site of the First International Satellite Land Surface Climatology Project Field Experiment (FIFE) in Kansas and one of the study sites of the Anglo-Brazilian Amazonian Climate Observational Study (ABRACOS) near the city of Manaus, Amazonas, Brazil. These two sites were selected because of the ready availability of relevant field data to validate and initiate the BATS-ABL model. The results of these tests are given a Masters thesis (Arain, 1994), and reported in two papers (Arain *et al.*, 1996; 1997).

Objective 4: Progress far exceeded original expectations not only in reconciling BATS land covers with those that can be recognized from space, but also in then applying remotely-sensed land cover data to map aggregate values of BATS parameters for heterogeneous covers and interpreting these parameters in terms of surface-atmosphere exchanges. The results are reported in a paper (Arain *et al.*, 1997).

Because the rate of progress under this contract exceeded that anticipated in the original proposal and on the basis of insight gained in the course of the research, it proved possible and appropriate

to expand the scope of activity. Additional research and development tasks were undertaken, as follows.

A. Provision of a State-of-the-Art Review

Contributions to a state-of-the-art review were made in the form of a paper reviewing current understanding of soil-atmosphere-vegetation interactions (Shuttleworth, 1995) and in the form of a paper expressing opinion on the need for, and form of, simplifying assumptions for surface-atmosphere exchanges (Shuttleworth, 1996). In addition, an international workshop was organized which focused on reviewing the state-of-the-art of research relevant to the contract. This workshop took place in February 1994 in Tucson, Arizona and has since become known as The Tucson Aggregation Workshop. With NASA approval, convening, organizing, and hosting of this workshop and then subsequently editing the conference proceedings into a Special Issue of the *Journal of Hydrology* (Michaud and Shuttleworth, 1997) was sponsored under this grant.

B. Investigation of the Impact of Spatial Variability in Precipitation

Modeling results at the FIFE and ABRACOS sites demonstrated the need for additional research to investigate whether the spatial variability of natural rainfall is significant in the aggregate representation provided by surface parameterization schemes (see second annual report). The results of this additional study were reported in a paper (White *et al.*, 1997).

C. Provision of Revised BATS Parameters from Field Data

In the course of our modeling studies at the ABRACOS site it became apparent that the standard vegetation parameters prescribed for use in BATS did not provide an ideal representation of measured surfaces exchanges and that revision was required, see Arain *et al.* (1997). There was particular interest in determining the adequacy of BATS when describing surface atmosphere interactions of heterogeneous vegetation cover in the semi-arid southwestern U.S., and field studies were carried out to provide relevant values of vegetation parameters. Although such studies have drawn the majority of their financial support from elsewhere, some partial student support was provided under this grant. The results of this study were reported in a paper (Unland *et al.*, 1996).

2. Summary of Major Findings

The major scientific findings of this study can be summarized as follows.

(a) Model tests with the BATS-ABL model initiated and validated with data from the FIFE site (Arain, 1994; Arain *et al.*, 1996) and the ABRACOS site (Arain *et al.*, 1997) suggest that simple aggregation rules for combining the values of vegetation parameter in a heterogeneous landscape calculate area-average fluxes that are acceptably similar to those calculated with explicit representation of separate vegetation patches under a wide range of hydrometeorological conditions and for a range of surface covers appropriate to these two sites.

- (b) Shortcomings in the representation of heterogeneous vegetation cover using aggregation rules are possible for landscapes and climates in which there are either large differences in soil moisture due to artificial irrigation providing wet and dry soil covering approximately equal areas (Arain *et al.*, 1996); or where there are approximately equal areas of shallow-rooted and deep-rooted vegetation growing in a climate where the shallow-rooted vegetation can come under soil moisture stress but the deep-rooted vegetation does not (Arain *et al.*, 1997).
- (c) When aggregation rules are applied across the conterminous U.S. using the USGS/EDC prototype land cover data to calculate area-average values of BATS vegetation parameters for $2.8^\circ \times 2.8^\circ$ and $1^\circ \times 1^\circ$ grid meshes, the calculated values of key vegetation parameters (aerodynamic roughness, minimum stomatal resistance, maximum and minimum leaf area indices and albedo) are significantly different from those for the most common cover in each grid square, the latter values being those that would normally be used in a General Circulation Model (Arain *et al.*, 1997).
- (d) Stand-alone BATS simulations with vegetation parameters corresponding to aggregate-cover and most-common cover, and using forcing variables from the ISLSCP CDROM-1 for $1^\circ \times 1^\circ$ grid squares across the conterminous U.S., gave surface fluxes which agreed with each other better than might have been anticipated because of mutual cancellation between the differences in more than one vegetation parameter. However there were sizable regions of the U.S. where significant difference occurred (Arain *et al.*, 1997).
- (e) The "standard" set of soil and vegetation parameters prescribed within BATS for semi-arid land cover (Unland *et al.*, 1996) and for tropical land covers (Arain *et al.*, 1997) does not provide adequate simulation of measured surface fluxes; however, revised parameters can be defined to remove this weakness.
- (f) Even in a semi-arid environment which is subject to locally heavy, convective rain storms, the heterogeneity in natural, rain-induced soil moisture patterns does not give rise to significant differences between the true area-average energy fluxes and those calculated assuming a single-point aggregate description. This is because the amount of rain falling in an individual precipitation event is a comparatively small proportion of the total amount of water that can be stored within the rooting zone of vegetation. It is likely that this result is more generally true in other environments.
- (g) The state-of-the-art review of research into the aggregate description of land-atmosphere interactions for heterogeneous surfaces (Shuttleworth, 1995; Shuttleworth, 1996; Michaud and Shuttleworth, 1996) concluded the following.
- (i) Substantial progress has now been made in producing aggregated representations of flat terrain and, in most studies thus far, simple aggregation rules applied to surface properties give simulated surface energy fluxes that are within 10% of fluxes produced from models with full representation of heterogeneity.

(ii) Aggregation rules are relatively straightforward in the case of patch-scale heterogeneity of vegetation-dependent controls on surface exchanges, but aggregation of soil hydraulic properties and soil moisture remains problematic. Additionally, some of the effects of mesoscale heterogeneity in surface cover on atmospheric turbulence and convection will need to be addressed through more complicated types of parameterization.

(iii) There is convincing evidence that the regional energy balance is insensitive to mild to moderate topography provided that surface vegetation and water availability are uniform. However, in mountainous terrain, the influence of topography on near-surface meteorology and vegetation cover will most likely also need to be considered.

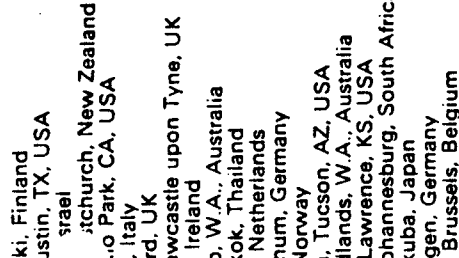
(iv) The values of simple combinations of remotely sensed radiances representing area-average measurements are only slightly influenced by subpixel variability, although the averaging of derived variables based on these radiances offers a much greater challenge, especially with sparse canopies.

3. List of Publications Resulting from this Contract

1. Arain, A.M., 1994: Spatial aggregation of vegetation parameters in a coupled land surface-atmosphere model, MS Thesis, Department of Hydrology and Water Resources, University of Arizona, Tucson, 150p.
2. Arain, A.M., J. Michaud, W.J. Shuttleworth, and A.J. Dolman, 1996: Testing of vegetation parameter aggregation rules applicable to the Biosphere-Atmosphere Transfer Scheme (BATS) and the FIFE site, *Journal of Hydrology*, 177, 1-22.
3. Arain, A.M., J.M. Michaud, W.J. Shuttleworth, Z-L Yang and A.J. Dolman, 1997: Mapping surface cover parameters using aggregation rules and remotely-sensed cover classes, Submitted to *Quarterly Journal of the Royal Meteorological Society*.
4. Michaud, J.D. and W.J. Shuttleworth, 1997: Executive Summary of the Tucson Aggregation Workshop, *Journal of Hydrology*, (in press).
5. Shuttleworth, W.J., 1995: Soil-vegetation-atmosphere relations: process and prospect, in *The Role of Water and the Hydrological Cycle in Global Change*, (H.R. Oliver, ed.), NATO ASI Series, Vol. I 31, Springer-Verlag, Berlin Heidelberg.
6. Shuttleworth, W.J., 1996: Hydrological models, regional evaporation and remote sensing: let's start simple and maintain perspective, in *Global Environmental Change and Land Surface Processes in Hydrology: The Trials and Tribulations of Modeling and Measurement*, (S. Sorooshian and V.H. Gupta, eds.), Springer-Verlag, New York, (in press).

7. Unland, H.E., P.R. Houser, Z.-L. Yang, and W.J. Shuttleworth, 1996: Surface flux measurement and modeling at a semi-arid Sonoran desert site, *Agricultural and Forest Meteorology*, (in press).
 8. White, C.B., P.R. Houser, A.M. Arain, Z.-L. Yang, and W.J. Shuttleworth, 1997: The aggregate description of semi-arid vegetation with precipitation-generated soil moisture heterogeneity, Submitted to *Research in Hydrology and Earth Systems Sciences*.
- Papers in Special Issue of Journal of Hydrology which were edited under this contract.*
9. Dolman A.J. and E.M. Blyth, 1997: Patch scale aggregation of heterogeneous land surface cover for mesoscale models, *Journal of Hydrology*, (in press).
 10. Kabat P., R.W.A. Hutjes, and R.A. Feddes, 1997: Scaling Soil Properties at field and catchment scale, *Journal of Hydrology*, (in press).
 11. Moran, M.S., K.S. Humes, and P.J. Pinter Jr., 1997: The scaling characteristics of remotely sensed variables for sparsely-vegetated heterogeneous landscapes, *Journal of Hydrology*, (in press).
 12. Noilhan, J, P. Lacarre, H. Dolman, and E. Blyth, 1997: Defining area-average parameters in meteorological models for landsurfaces with mesoscale heterogeneity, 1997, *Journal of Hydrology*, (in press).
 13. Pielke R.A., X. Zeng, T.J. Lee, and G.A. Dalu, 1997: Mesoscale fluxes over flat landscapes for use in larger scale models, *Journal of Hydrology*, (in press).
 14. Raupach, M.M. and J.J. Finnigan, 1997: The influence of topography on meteorological variables and surface-atmosphere interactions, *Journal of Hydrology*, (in press).
 15. Sellers, P.J., M.D. Heiser, F.G. Hall, S.B. Verma, R.L. Desjardins, P.M. Schuepp, and J.I. MacPherson, 1997: The impact of using area-averaged land surface properties--topography, vegetation condition, soil wetness--in calculations of intermediate scale ($\sim 10 \text{ km}^2$) surface-atmosphere heat and moisture fluxes, *Journal of Hydrology*, (in press).
 16. Thorton, P.E. and S.W. Running, 1997: Mapping daily surface meteorological variables in complex terrain with MTCLIM-3D, *Journal of Hydrology*, (in press).
 17. Wood, E.F., 1997: Effects of soil moisture aggregation on surface evaporative fluxes, *Journal of Hydrology*, (in press).

(Copies of papers 2-17 are provided herewith).



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A realistic model of surface-atmosphere exchanges was created by coupling the Biosphere-Atmosphere Transfer Scheme (BATS) with an advanced, two-dimensional model of the atmospheric boundary layer. This was initiated and tested using data obtained from the First International Satellite Land Surface Climatology Project Field Experiment (FIFE). This model was used to investigate the acceptability of simple rules for defining the aggregate value of the parameters required to specify surface interactions, as applied to heterogeneous mixes of vegetation types allowed in BATS but appropriate to the FIFE site, namely short and long grass, mixed crops, and irrigated crops. Under the range of meteorological and surface conditions relevant to FIFE and as used in this study, these rules are shown to estimate aggregate parameters which give surface fluxes similar to those calculated with an explicit representation of separate vegetation patches, except in the particular case of artificially wetted (irrigated) patches set in an otherwise dry landscape.

Increasingly realistic and complex submodels, such as the Biosphere-Atmosphere Transfer Scheme (BATS; Dickinson et al., 1986), are routinely used in general circulation models (GCMs) to describe land surface processes. However, even with these advanced land surface parameterizations, the land surface and near-surface

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meteorology are considered spatially uniform within GCM grid elements that range in size, currently from 10^4 – 10^6 km². For each gridpoint, a single set of vertical calculations is made to determine energy fluxes and surface states, such as temperature and soil moisture content. A number of parameters are used to describe the soil and vegetation characteristics at the gridpoint. Values of these parameters are usually selected to correspond to the most common vegetation type (called 'cover' here) and soil type within the grid square. It is widely recognized that failure to address subgrid heterogeneity affects the area-average surface fluxes and may be a significant limitation of GCMs. Two general solutions have been proposed for addressing subgrid variability: first, using 'effective' or 'aggregate' parameters, and second, addressing heterogeneity in a statistical sense (see, for example, Famiglietti and Wood, 1991).

In the aggregate approach, the single, vertical description of the land surface exchange processes within each grid square is retained, but parameters are estimated so that they provide a reasonably realistic description of area-average surface exchanges of energy, water vapor, and momentum (e.g. Shuttleworth, 1991). It is anticipated that the value of these so-called 'aggregate parameters' will be derived using a suitable averaging procedure from information on the relative proportion of different vegetation covers in each grid element and prior estimates of the parameter values for each cover type present. The nature of the averaging procedure should follow from the physical nature of the exchanges which each parameter influence.

Aggregation studies are typically conducted by comparing fluxes computed with a model describing heterogeneous surfaces to fluxes computed with a model describing the equivalent spatially homogeneous model. In some situations, the performance of the model has been checked against and initiated using field data (e.g. Noilhan and Lacarrère, 1995), but this has not been a universal practice so that, at this writing, the data from large-scale field experiments (HAPEX-MOBILHY, André et al., 1990; FIFE, Sellers et al., 1992; HAPEX-Sahel, Goutorbe et al., 1994) remain an under-used resource. Several modeling experiments have examined heterogeneity at length scales intermediate between 1 m and 10 km (Mason, 1988; Wood and Mason, 1991; Blyth et al., 1993; Guo and Schuepp, 1994). These studies, which used fairly simple land surface models, provided support for the aggregation approach when land surface variability creates near-surface atmospheric perturbations that do not penetrate through the entire thickness of the planetary boundary layer (Shuttleworth, 1988). Other experiments conducted at larger scales (where perturbations can potentially affect the entire boundary layer) also have provided support for the aggregation approach (Avisar and Pielke, 1989; Lhomme et al., 1994; Noilhan and Lacarrère, 1995).

This paper addresses the computationally efficient aggregation approach and focuses on defining aggregate parameters that describe vegetation characteristics. An advanced and widely studied soil-vegetation-atmosphere model (the Biosphere Atmosphere Transfer Scheme (BATS)) was used to explore and examine aggregation rules with intermediate scale (1 km) vegetation heterogeneity. It is a refinement of previous coupled studies at intermediate scales because it uses an advanced land surface model and observed field data. The general philosophy of this investigation

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was to test various aggregation strategies under realistic conditions. The following guidelines were intended to produce maximum realism in the simulations:

- (a) Select components for the coupled land-atmosphere model which are accepted and realistic: Mason's two-dimensional model of the atmospheric boundary layer (hereafter called ABL model, Mason and Sykes, 1980) and BATS (Dickinson et al., 1986);
- (b) Conduct the analysis using data from a previous large-scale field experiment at times corresponding to periods of intensive study when observed data are available for model initiation and validation: The FIFE experimental site in Kansas, which represents mid-latitude prairie, was selected;
- (c) Investigate the aggregation rules only for vegetation covers defined for BATS and for covers that might realistically be expected to occur in the general vicinity of the experimental site; and
- (d) Examine a variety of field conditions which are considered representative of those observed during the relevant field experiment.

The coupled BATS-ABL model was used to calculate the difference between area-average and aggregate land surface parameterizations under a range of simulated surface covers but using observed initial atmospheric profiles. The BATS-ABL model is described in Section 2; the use of FIFE data is described in Section 3; and various model sensitivity studies are included in Section 4. The results of the aggregation studies are given in Section 5 and discussed in Section 6.

2. The BATS-ABL coupled model

2.1. Atmospheric boundary layer component

The ABL model creates two-dimensional fields of wind velocity, pressure, temperature, and humidity, which are in equilibrium with a heterogeneous land surface once the model has been integrated to steady state. It is a first-order closure model, and the turbulent mixing is assumed to be the same for momentum, heat, and water vapor. The equations used are the two-dimensional form of the ensemble-averaged, Boussinesq approximation to the Navier-Stokes equations, the continuity equation for an incompressible atmosphere, and the conservation equations for humidity and heat. A mixing length formulation is used to derive the Reynolds stress. In neutral and unstable boundary layers, which is the normal situation in this study, the mixing length increases with height up to a maximum of one-third of the boundary layer depth (0.33 km).

The lower boundary conditions of the ABL model are a no-slip condition for wind and values of sensible heat flux and surface humidity which are determined by the land surface model. The upper boundary conditions are no flux of heat and moisture at the top of the boundary layer, i.e. no mixing in the vertical direction at this height. This somewhat artificial condition has little consequence, because the model is usually run forward in time for short periods to allow the lower boundary layer to come into

equilibrium with the surface. In this study, the ABL model has 32 horizontal grid-points, each 62.5 m apart, and 16 vertical gridpoints (logarithmic spacing with five layers in the first 10 m). The interface height between the atmospheric and the land surface submodels was set to 2 m above the ground because this was the height at which temperature and humidity were measured at the FIFE site. The model has periodic boundary conditions, with the variables at the one end of the grid after each time step re-assigned to the other end of the grid for the next time step. The integrating time of the model depended on the amount of surface variation and the spatial dimension of the variation, but typically corresponded to 6–10 min. For additional details of ABL model, the reader is referred to Mason (1988), Wood and Mason (1991), Blyth et al. (1993), and Arain (1994).

2.2. Biosphere–Atmosphere Transfer Scheme

BATS is a comprehensive soil–vegetation–atmosphere–transfer scheme which is based on current understanding of biological processes at plot scale. A schematic representation of processes simulated by BATS is given in Fig. 1. BATS uses separate model components for its radiative and hydrological parts. The soil is divided into three layers for water balance calculations, and the model recognizes the separate contributions of soil and vegetation towards surface fluxes. BATS calculates latent and sensible heat fluxes using the classical equations in which heat flux is proportional to the temperature gradient between the surface and the air, and latent flux is proportional to the humidity gradient. Ground heat flux is calculated as a residual from the surface energy balance, and soil temperature is calculated using the force–restore method (Deardorff, 1978; Dickinson, 1988).

Two transfer phases are considered in the movement of water and heat flux from the canopy to the atmosphere. First, the exchange with the air within the canopy and, second, the exchange between the canopy air and the atmosphere overlying the canopy. The processes considered within the canopy are sensible and latent heat exchange with the atmosphere, the absorption of solar radiation, and the presence of canopy surface moisture as a result of dew or rainfall. Over dry areas of the canopy, stomatal resistance controls the escape of water from saturated conditions within the leaf to the adjacent external air. Stomatal resistance increases as soil moisture decreases; details of this parameterization are given in Appendix A and Section 5.1.

All surface fluxes are proportional to the drag coefficient C_D , which is a function of the interface height between BATS and the atmospheric model (see Appendix B for details). As will be shown in Section 4.3, sensitivity to the interface height could be an issue with the BATS–ABL model, and this, we believe, is quite likely to be a more general but previously unrecognized feature of coupled models.

3. Model parameterization, and forcing for the FIFE site

The First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) was a multiscale, multidisciplinary experiment conducted on a

crop were selected for the aggregation studies. The parameter values associated with the selected land covers are given in Table 1.

Soil texture class 7 and soil color class 5 were selected as being appropriate for the FIFE site based on observations made by Famiglietti et al. (1992), Stewart and Verma (1992), Ungar et al. (1992), and with advice from Greg Colello (Personal communication, 1994). Soil parameter values for these classes are given in Table 2.

The BATS-ABL model was initialized and forced with observed atmospheric and ground data from the FIFE site for specific days and times. The quality and availability of data was a primary consideration in the selection of days for which model experiments were conducted, with the time of day for the experiments selected to correspond to the time of radiosonde measurements. Days were selected in order to sample the seasonal variation in fluxes, soil moisture conditions, and the phase of vegetation greenness. The required data were obtained from the FIFE Information System. The radiosonde data were collected by Brutsaert and Sugita (1992), while the ground data were originally collected by various investigators at several FIFE sites (Kanemasu et al., 1992; Sellers et al., 1992). Betts and Ball (1993) subsequently performed quality control on these data and averaged them in space (to give a site average) and time (to give a 30-min average).

The ABL model was initialized with observed radiosonde profiles of potential temperature, humidity, wind speed (u and v components), and pressure, supplemented with ground measurements of the same variables. The initial profiles were assumed to be spatially uniform. It was necessary to initialize nine state variables in BATS (mostly soil temperature and moisture; see Table 3). Because the BATS-ABL model is run forward for only a few minutes, the model's fluxes are very sensitive to initialization, and the initial state variables were therefore obtained by running a 'stand-alone' version of BATS to generate the required state variables at the dates and

Table 1
Values of the BATS vegetation parameters for the FIFE-relevant cover types

Parameter	Mixed crop	Short grass	Long grass	Irrigated crop
Maximum fractional vegetation cover	0.85	0.8	0.8	0.8
Factor for seasonal change in vegetation cover	0.6	0.1	0.3	0.6
Roughness length (m)	0.06	0.02	0.1	0.06
Depth of rooting zone (m)	1	1	1	1
Depth of the upper soil layer (m)	0.1	0.1	0.1	0.1
Fraction of water extracted by upper layer roots	0.3	0.8	0.8	0.3
Albedo for wavelength $> 0.7 \mu\text{m}$	0.1	0.1	0.08	0.08
Albedo for wavelength $< 0.7 \mu\text{m}$	0.3	0.3	0.3	0.28
Minimum stomatal resistance (s m^{-1})	120	200	200	200
Maximum leaf area index	6	2	6	6
Minimum leaf area index	0.5	0.5	0.5	0.5
Stem and dead matter area index	0.5	4	2	2
Inverse square root of leaf dimension ($\text{m}^{-1/2}$)	10	5	5	5
Light sensitivity factor ($\text{m}^2 \text{W}^{-1}$)	0.02	0.02	0.02	0.02
Zero plane displacement	0	0	0	0

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Table 2

BATS soil parameters used in this study

(a) Soil texture parameters for class 7

Parameter	Value
Porosity	0.51
Minimum soil suction (mm)	200
Saturated hydraulic conductivity (mm s^{-1})	0.0045
Ratio of saturated thermal conductivity of that to loam	0.95
Exponent B	6.8
Moisture content relative to saturation at which transpiration ceases	0.378

(b) Albedo for soil color class 5

Condition	Albedo
Dry soil, wavelength $< 0.7 \mu\text{m}$	0.16
Dry soil, wavelength $> 0.7 \mu\text{m}$	0.32
Wet soil, wavelength $< 0.7 \mu\text{m}$	0.08
Wet soil, wavelength $> 0.7 \mu\text{m}$	0.16

times of interest. Four such stand-alone simulations were performed to obtain the initial state variables relevant to each cover type used in the study. The stand-alone model was run from May 26, 1987, to October 11, 1987, using observed forcing variables at 30-min intervals (see Table 3 for forcing variables).

4. Model tests

A variety of tests were performed prior to (and during) the aggregation simulations to determine if the model was operating correctly. These are over viewed in the following sections.

4.1. Accuracy of calculated fluxes

Simulated fluxes from the stand-alone version of BATS were compared to observed fluxes obtained from Betts and Ball (1993). Both the observed forcing variables and the observed fluxes are site averages. Tall grass cover type, which corresponds most closely to the actual vegetation at the FIFE site, was used for this comparison. The stand-alone BATS simulation ran for 143 days in the summer of 1987, and calculated fluxes were compared to observed fluxes for the 57 days for which flux measurements are available. The mean difference between simulated and observed fluxes was 6 W m^{-2} for latent heat (with simulated evaporation too small), 6 W m^{-2} for sensible heat (model heating of the air too small), and 15 W m^{-2} for ground heat flux (model heating of the ground was too great). At a 30-min time step, the correlation coefficient between simulated and observed fluxes was 0.80 for latent heat, 0.80 for sensible heat,

Long grass

Irrigated crop

0.8	0.8
0.3	0.6
0.1	0.06
1	1
0.1	0.1
0.8	0.3
0.08	0.08
0.3	0.28
200	200
6	6
0.5	0.5
2	2
5	5
0.02	0.02
0	0

Table 3
BATS state variables, driving variables, and parameters

	Symbols in text
<i>States</i>	
Surface soil temperature ($^{\circ}\text{K}$)	T_{s1}
Subsurface temperature at depth of 20 cm ($^{\circ}\text{K}$)	T_{s2}
Temperature of foliage ($^{\circ}\text{K}$)	T_f
Temperature of air within foliage ($^{\circ}\text{K}$)	T_{af}
Water in upper soil layer (mm)	
Water in root zone soil (mm)	
Total water in soil (mm)	
Ground wetness factor (fraction)	
Dew on leaves (mm)	
Sea ice fraction (not used)	
Snow water equivalent (not used)	
Snow age (not used)	
<i>Driving variables (boundary inputs)</i>	
Incident solar radiation (W m^{-2})	V_a
Wind velocity (m s^{-1})	T_a
Air temperature at interface level ($^{\circ}\text{K}$)	Q_a
Specific humidity at interface level (g Kg^{-1})	
Precipitation (mm)	P_{surf}
Surface pressure (mbar)	
Downward longwave radiation (W m^{-2})	
<i>Parameters</i>	
Interface height (m)	z_1
Vegetation parameters (see Table 1)	
Land or sea flag	
Plant carbon uptake ($\text{Kg of C m}^{-2} \text{s}^{-1}$)	
Plant respiration rate ($\text{Kg of C m}^{-2} \text{s}^{-1}$)	
Soil texture parameters (see Table 2)	
Soil color parameters (see Table 2)	

0.81 for ground heat flux, and 0.99 for net radiation. While the observed and simulated fluxes did not match exactly, the simulated fluxes were sufficiently similar to measured values to allow a measure of confidence in BATS as a reasonable description of the land surface.

4.2. Integration to equilibrium

When the coupled model is run forward from the observed initial conditions, there is a short period during which the land model and atmospheric model are adjusting to each other. The evolution of fluxes and states was examined to determine length of time to reach equilibrium; typically this took about 400 s. Unless otherwise stated, model results presented herein are for an integration time of 600 s.

4.3. Interface height sensitivity tests

As mentioned in Section 2, the calculation of surface energy fluxes may be sensitive to the arbitrary selection of the interface height between the two submodels in BATS-ABL. To evaluate this possibility, integrations were carried out with the BATS-ABL model using two very different interface heights, namely 2 m (the standard height used in this study) and 20 m (the standard height of tall vegetation in BATS). Fig. 2 illustrates the initial observed profiles over short grass for noon on August 15, 1987, and the ensuing atmospheric profiles calculated by BATS-ABL after running for 12 s and 10 min. BATS-ABL smooths the profile and 'retunes' simulated atmospheric profiles so that they are in equilibrium with the surface. It appears that BATS provides a more rapid analytic integration of the variations between 2 and 20 m than the finite difference integration used in ABL and, in consequence, the

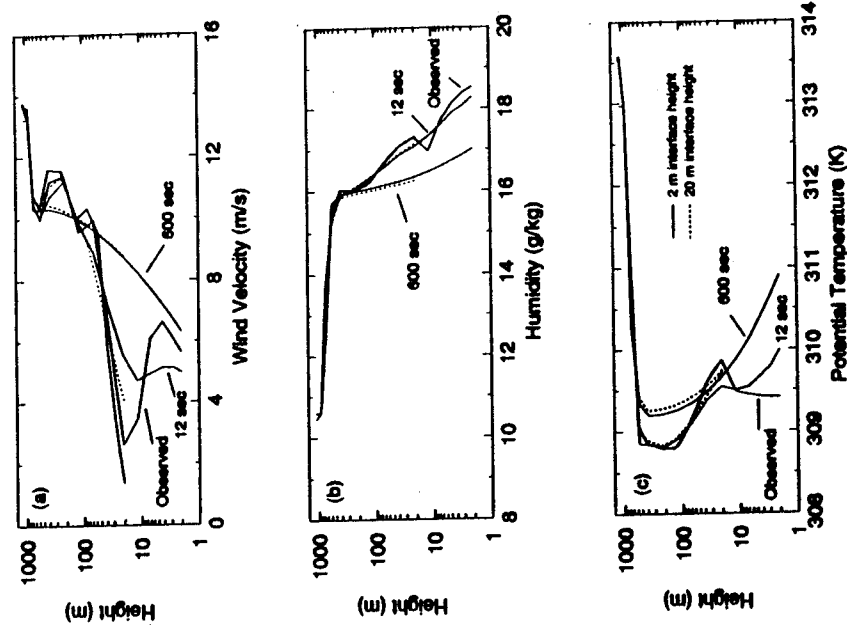


Fig. 2. Profiles of (a) wind, (b) humidity, and (c) temperatures at noon on August 15, 1987. Initial observed profiles are shown, along with simulated profiles for short grass. Simulated profiles were generated by BATS-ABL after running for 12 s and 600 s with the interface height set to 2 m and 20 m.

While the observed and
xes were sufficiently similar
ATS as a reasonable descrip-

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e of 600 s.

final atmospheric profile after the initial adjustment remains systematically different for the two interface heights.

In order to remove this initialization feature (a possible cause of different surface fluxes between the two models) both the 2-m and 20-m versions of BATS–ABL were next initiated with the same smooth atmospheric profiles (calculated by running the 2-m version of the model for 10 min after initialization with observed profiles). When the models were initiated with the same smooth profiles, the evolution of the last-mentioned model for 20 min, the fluxes of sensible heat, latent heat, and soil heat were 220, 325, and 38 W m^{-2} , respectively, for the 2-m interface, and 234, 282, and 60 W m^{-2} , respectively, for the 20-m interface. Hence, the difference between ABL's and BATS' representation of near-surface aerodynamic transfer is of some consequence to the calculated surface fluxes, in spite of initializing with near-surface atmosphere profiles. The specification of near-surface turbulent transport is an inherent property of the land surface scheme for which investigation of aggregate surface properties are made; in this case, they are a characteristic of BATS. Clearly, studies such as this one must make proper recognition of the possibility that calculated values of the surface fluxes may show sensitivity to the selected height of the interface between the atmospheric portion of the coupled model and the land surface scheme under test. In the case, these differences were shown not to be significant in the context of testing hypothetical aggregation rules (see below). For greater detail, the reader is referred to Arain (1994).

In order to observe the difference between area-average fluxes from the two-patch model and fluxes from a corresponding aggregate cover for different model interface heights (2, 10, 15, 20, 35, 40, and 50 m above the ground), a series of integrations were made with BATS–ABL applied to two adjacent patches of 'short grass' and 'long grass'. The size of each patch was 1 km, and both regions were initialized with states from off-line BATS run. The atmospheric profiles were initialized to those observed at the FIFE site at noon on August 15, 1987. Area-average and aggregate fluxes were computed after integration for 20 min.

Figs. 3(a) to (c) illustrate the variations of area-average sensible heat, latent heat, and soil heat with the assumed height of the interface. The height dependence of the individual fluxes is shown to be significant and of the order of several $10\text{s of } \text{W m}^{-2}$. However, the criterion important to the present study, namely the difference between the calculated area-average fluxes with patches of vegetation and aggregate vegetation, shows insignificant variation with height, typically on the order of $1\text{--}2 \text{ W m}^{-2}$ (see Figs. 3(d)–(f)). On this basis, we concluded that the sensitivity of the model to the interface height does not pose a problem for the current study.

4.4. Effect of model domain and patch size

The size of domain represented in numerical, time-dependent models is significant, because it not only affects the averaging schemes, but also the computational time, while the spatial discretization of a finite difference scheme can affect stability and accuracy of the solution. The selection of patch and domain size also depends on the

is systematically different because of different surface is of BATS-ABL were calculated by running the observed profiles). When the evolution of the 2-m n. After running the last- nent heat, and soil heat were face, and 234, 282, and difference between ABL's transfer is of some conse- alizing with near-surface urbulent transport is an nvestigation of aggregate teristic of BATS. Clearly, the possibility that calcu- the selected height of the model and the land surface not to be significant in the w). For greater detail, the fluxes from the two-patch r different model interfaces series of integrations were of 'short grass' and 'long were initialized with states alized to those observed at d aggregate fluxes were

sensible heat, latent heat, e height dependence of the r of several $10\text{s of } \text{W m}^{-2}$. nely the difference between tion and aggregate vegeta- on the order of $1\text{--}2 \text{ W m}^{-2}$ isitivity of the model to the tudy.

ndent models is significant, o the computational time, me can affect stability and in size also depends on the

natural heterogeneity of the area under consideration. A series of experiments were conducted using two patches of mixed crop and short grass, with the model initiated to environmental conditions observed at the FIFE site at noon on August 15, 1987. The size of domain and patch tested for each cover type were (a) 1-km domain, with two $\frac{1}{2}$ -km patches, (b) 2-km domain, with four alternate $\frac{1}{2}$ -km patches, (c) 2-km

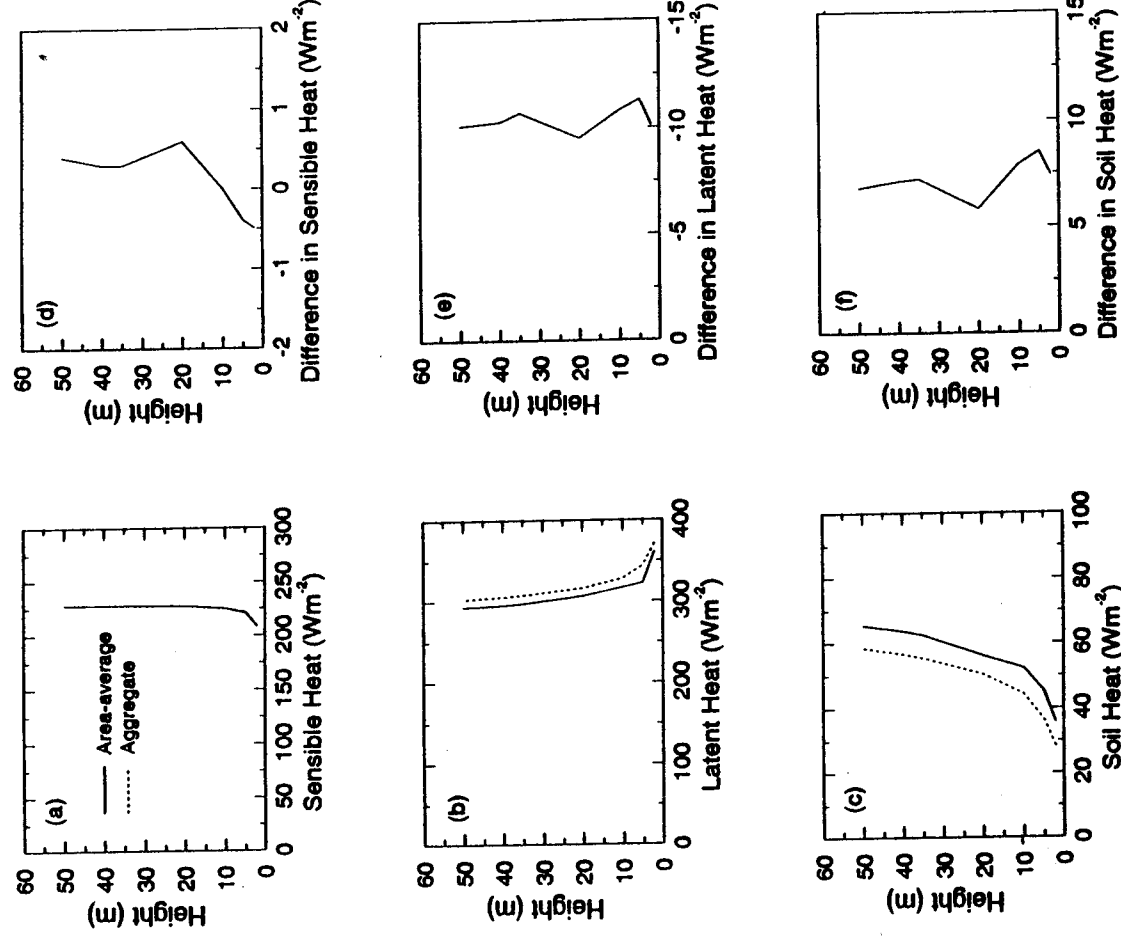


Fig. 3. Dependence of simulated fluxes on the interface height between the BATS and ABL models: (a), (b), and (c) represent the area-average fluxes calculated for landscape comprised of short grass and long grass patches and the equivalent fluxes from a homogenous landscape of 'aggregate' vegetation; (d), (e), and (f) show the difference between area-average and aggregate fluxes.

domain, with two 1-km patches, (d) 4-km domain, with four alternate 1-km patches, and (e) 4-km domain, with two 2-km patches.

Integrations with different domain and patch sizes were run for equal time periods (10 min). Results showed that patch and domain size make little difference in calculated surface fluxes. On the basis of these results, 1-km patches in a 2-km domain were selected as the standard for the remainder of the aggregation tests.

4.5. Heterogeneity of fluxes

The effect of cover type on fluxes is illustrated in Fig. 4. This four-patch simulation is for a time of vigorous fluxes (noon on August 15). The decreased evaporation at the left edge of Patch 2 is due to advection of relatively moist cool air from Patch 1 (the wind is from the left). The decrease in latent heat at the left edge of Patch 2 is balanced by increases in sensible heat and ground heat flux. The perturbation at the left edge of Patch 2 is cancelled (in terms of area averages) by an increase in evaporation at the left edge of Patch 3. The fact that stomatal resistance used in BATS has an assumed

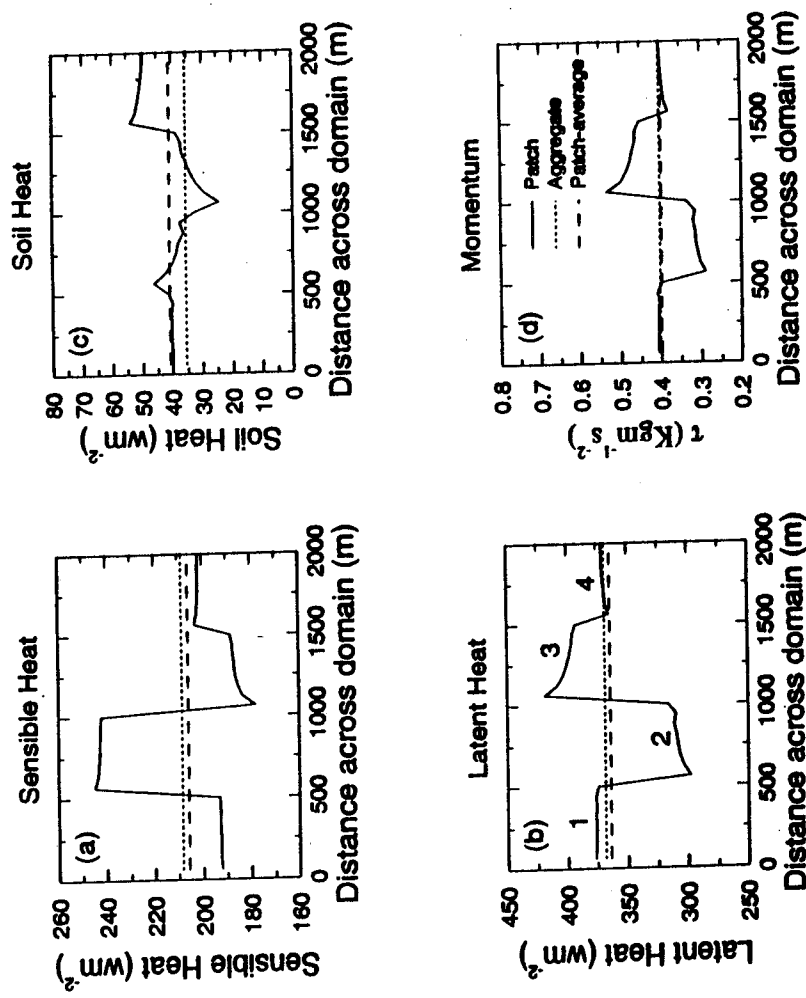
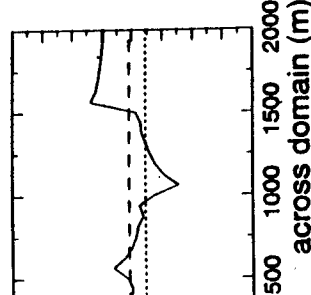


Fig. 4. Fluxes across a domain consisting of four patches of mixed crop (1), long grass (2), short grass (3), and irrigated crop (4). Simulations are for noon on August 15, 1987.

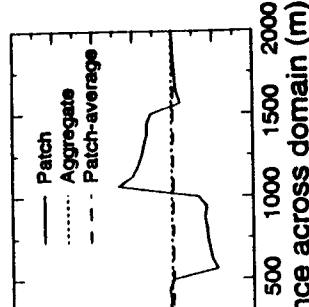
for equal time periods
little difference in calcu-
lations in a 2-km domain were
on tests.

This four-patch simulation
increased evaporation at the
cool air from Patch 1 (the
edge of Patch 2 is balanced
urbane at the left edge of
use in evaporation at the left
in BATS has an assumed

Soil Heat



Momentum



(1), long grass (2), short grass (3).

dependency on meteorological variables does not seem to compromise the effectiveness of this cancellation significantly.

5. Definition and tests of aggregation rules

5.1. Aggregation rules for BATS parameters

The purpose of an aggregation rule is to define an "effective parameter which is the weighted average of component covers, through that function involving the parameter which most nearly links it to a linear average of the associated flux" (Shuttleworth, 1991). Averaging rules for most parameters are thus based on the manner in which they influence fluxes. For example, because solar radiation is related linearly to albedo, it follows that albedo should be aggregated as an area-weighted arithmetic average. The aggregation rules used in this study are defined in Table 4. Area-weighted arithmetic averaging was performed for all parameters except aerodynamic roughness length, z_0 , and stomatal resistance, r_s .

Roughness length and stomatal resistance are of primary importance in determining surface fluxes. Roughness is averaged in such a way that the effective value equals that value for which a homogeneous area would provide a surface stress equal to the spatial average of a heterogeneous area. The averaging is based on the blending height (Wierenga, 1986; Mason, 1988). The value of blending height (in m) can be calculated by the equation given in Table 4 or approximated by the patch length (in m) divided by 200 (Mason, 1988).

Leaf stomata are the primary route for the flow of water vapor from the near-saturated surface conditions within the leaf to the canopy air outside. In BATS, a minimum stomatal resistance, $r_{s\min}$, is attenuated using multiplicative stress functions

Table 4

Aggregation rules for BATS vegetation parameters

Parameter	Aggregation rule	Comment
Maximum leaf area index (LAI MAX)	$LAI MAX = \sum w_i LAI MAX_i$	Linear effect on radiation and evaporation
Minimum stomatal resistance ($r_{s\min}$)	$\frac{1}{r_{s\min}} = \sum \frac{w_i}{r_{s\min i}}$	(i) Reciprocal rule: evaporation inversely related to stomatal resistance in BATS
	$\overline{r_{s\min}} = \sum w_i r_{s\min i}$	(ii) Linear rule
Roughness length for momentum (z_0)	$\ln \left(\frac{b}{z_0} \right)^2 = \sum w_i \ln \left(\frac{b}{z_{0i}} \right)^2$	Blending height given as $b = 2(u^*/V_s)^{-1} l_c$ where l_c is the horizontal length scale

The subscript "i" represents cover (vegetation) type and "w_i" represents the fractional area of that cover type. Parameters (see Table 3 for a list) that use a linear area-weighted average have been omitted for brevity, with the exception of maximum leaf area index, which is included as an example.

determined by incident radiation, root resistance and vapor pressure deficit, and for seasonal effects. It is currently not clear how the averaging of stomatal resistance should proceed. Strictly speaking, the actual stomatal resistance, r_s not $r_{s, \min}$, should be averaged in BATS. However, the values of the parameters which determine the stress functions are the same for all BATS vegetation types, therefore averaging $r_{s, \min}$ is equivalent. In practice, the numerical value of the required aggregate $r_{s, \min}$ is virtually the same for the two alternative (linear and reciprocal) averaging rules given in Table 4 with the classes of land cover relevant to the FIFE site (tall and short grass, crops, and irrigated crops). Currently, therefore, it is not clear whether a reciprocal average or a linear average is preferable for $r_{s, \min}$ in BATS.

Because initial states to be used in aggregation runs with BATS-ABL vary with cover type, it was necessary to aggregate initial states for the aggregate-patch runs. In fact, this would not be necessary in an actual GCM simulation for a homogeneous grid.

5.2. Aggregation tests

Testing of postulated aggregation rules for BATS parameters at the FIFE site comprised the following sequence for each of the representative days and times:

- (a) Appropriately initialize the BATS-ABL model (see Section 3), then integrate for 10 min across a 2-km domain with two 1-km patches of different vegetation selected from the FIFE-relevant BATS land cover classes;
- (b) Apply the postulated aggregation rules to calculate aggregate model parameters and the initial BATS state variables. Introduce these into BATS-ABL to produce a spatially homogeneous model representing the average of the two vegetation patches used in the first simulation and integrate this model for 10 min; and
- (c) Compare the area-average fluxes calculated by the two-patch model with those given by the homogeneous vegetation with aggregate parameters.

Accordingly, the aggregation rules described in Section 5.1. were tested using initialization data from the FIFE data set at three times (about 08:00, 12:00, and 16:00 h local time) on August 15, 1987, three times on October 11, 1987 (about 8:00, 12:00, and 16:00 h), and two times on July 7, 1987 (about 12:00 and 16:00 h; observed data were only available at two times on July 7). Five sets of simulations were conducted for each of these eight times. The five simulations correspond to the following cover combinations:

- (1) Short grass-long grass.
- (2) Irrigated crop-long grass.
- (3) Mixed crop-long grass.
- (4) Mixed crop-irrigated crop.
- (5) Mixed crop-short grass.

In this way, a total of 40 two-patch simulations and 40 corresponding simulations with aggregate vegetation were performed for the 3 test days. The results, which consist of a comparison of the area average and aggregate fluxes of sensible heat

the short grass/long grass case. For short grass, the upper and root soil moisture were 16.1 and 327.1 mm, respectively, while those for long grass were 15.5 and 314.1 mm, respectively. However, the correspondence between patch and aggregate vegetation runs was poor in the October 11 runs for combinations in which an artificially wet (irrigated) soil was adjacent to a naturally dry soil (BATS unrealistically assumes that irrigation continues past the end of the growing season). Specifically, with the noon and afternoon model initiations, there was a disagreement when irrigated crops were introduced into a landscape of either long grass or mixed crop.

In summary, the postulated, BATS-specific, simple aggregation rules worked well in the absence of strong patch-to-patch contrasts in soil moisture. This would be expected when the patch-to-patch differences in latent and sensible heat are small, but in fact the aggregation representation was successful, even when there were quite large ($50\text{--}100\text{ W m}^{-2}$) contrasts in sensible and latent heat between the two patches. The aggregate descriptions for ground heat flux were not significant because the patch-to-patch differences in ground heat flux were small for each of the simulations.

5.3. Effect of soil moisture contrast

In the integrations carried out on October 11, 1987, the irrigated crop transpired at a potential rate, while the vegetation in the dry patches (mixed crop and long grass) was close to the wilting point. Fig. 6 shows the plant wilt factor, W_{LT}^i , plotted against relative soil moisture. Because W_{LT}^i is a non-linear function of the soil moisture ratio, the average value of W_{LT}^i evaluated for a dry soil and W_{LT}^i evaluated for a wet soil is not necessarily the same as W_{LT}^i evaluated for the soil moisture average. The effect of this non-linearity is illustrated by the patches of mixed crop and irrigated crop for the simulation at noon on October 11. As shown in Table 5, soil moisture for the aggregate simulation is the average of soil moisture from the two patches. The latent heat flux from the aggregate simulation is, however, 41% higher than the average of latent heat flux from the two patches.

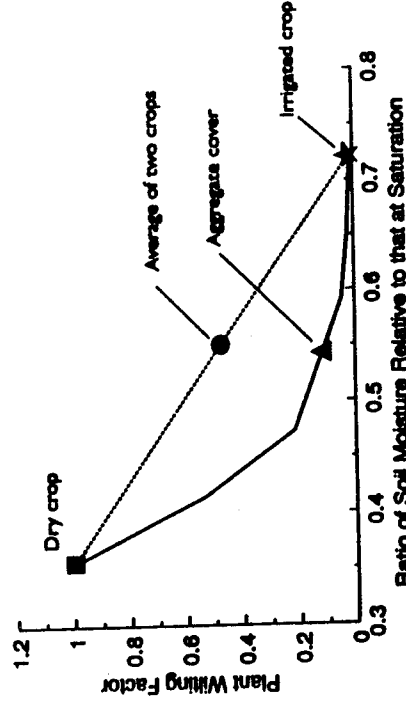


Fig. 6. Relationship of the plant wilt factor W_{LT}^i to soil moisture expressed as a fraction of saturation. Symbols denote the plant wilting factor for the simulations at noon on October 11, 1987.

and root soil moisture were 15.5 and 314.1 mm, and aggregate vegetation which an artificially wet unrealistically assumes that specifically, with the noon when irrigated crops were crop.

regation rules worked well moisture. This would be d sensible heat are small, even when there were quite between the two patches. ot significant because the or each of the simulations.

irrigated crop transpired at mixed crop and long grass) factor, W_{LT} , plotted against of the soil moisture ratio, evaluated for a wet soil is sture average. The effect of p and irrigated crop for the soil moisture for the aggregate patches. The latent heat r than the average of latent

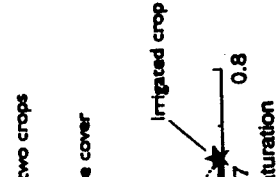


Table 5

Soil moisture and turbulent fluxes for October 11, 1987 (noon) simulation

	Mixed crop	Irrigated crop	Aggregate cover
Upper soil moisture	18 mm	36 mm	27 mm
Root zone soil moisture	220 mm	371 mm	295 mm
Latent heat flux	95 W m ⁻²	276 W m ⁻²	262 W m ⁻² (Patch average = 185)
Sensible heat flux	245 W m ⁻²	131 W m ⁻²	173 W m ⁻² (Patch average = 188)

The failure of aggregation in these simulations, which is caused by a non-linearity in root resistance, results in a real difference in the net turbulent fluxes returned to the atmosphere (374 W m^{-2} for the average of the two patches versus 435 W m^{-2} for the aggregate simulation). It should be noted that such a result would not be obtained for models which assume zero ground heat flux. In the case of the BATS model, it would be possible to avoid this aggregation problem by calculating the reciprocal average of root resistance, and then determining the aggregate soil moisture corresponding to that aggregate root resistance; however this is not an acceptable solution because doing so would break the water balance. There is a clear need to determine whether, in the real world, spatial differences in soil water availability can and do occur with sufficient regularity and with sufficient extent for this to be a significant issue in meteorological models. We return to this point in Section 6.

5.4. Alternate aggregation rules for stomatal resistance

In this study, several aggregation rules were examined for stomatal resistance, because previous studies have found support for both linear and reciprocal averaging rules, while Blyth et al. (1993) found that an average of linear and reciprocal averages gave the best results. However, in practice, there is little difference in the calculated surface energy balance irrespective of the selected aggregation rule. For noon on August 15, for example, there was a 2% difference between the latent heat obtained with the linear average and the latent heat obtained from the reciprocal average, despite the fact that the latent heat flux for individual patches differs by about 100 W m^{-2} (see Fig. 4). This difference was 1% for comparisons involving mixed crop and irrigated crop for noon on October 11. The results are similar for other fluxes and at other times. Therefore it appears that either the linear average or the reciprocal average (or the average of these two) works equally well in providing an estimate of the aggregate surface resistance in the case of the BATS land covers which are plausible in Kansas.

6. Summary and discussion

This paper provides additional support for the use of simple aggregation rules

expressed as a fraction of saturation. October 11, 1987.

which compute area-average vegetation parameters in land surface models. The primary contribution of this study is to demonstrate that simple aggregation rules, when applied to BATS parameters within a coupled model, were successful at producing simulations of area-average surface fluxes under a variety of realistic hydro-meteorological conditions at a mid-latitude prairie (the FIFE site). By successful, we mean that aggregate parameters gave rise to surface fluxes that are reasonably similar to those calculated as the area-average of separate vegetation patches. This success provides support for the viewpoint that the complicating effect of subgrid scale, near-surface advection (of heat, moisture, and momentum) may be insignificant in the context of defining effective, aggregate values for vegetation-related parameters.

It is important to recognize that the success of the vegetation parameter aggregation in this study is a necessary (but not sufficient) justification for adoption of the suggested aggregation rules. In particular, some important aggregation rules, the most important being for surface resistance, remain untested in this study. Therefore further research similar to this is required using different initiation data from different sites where there are strong contrasts in these unstudied parameters.

Another contribution of this study is to demonstrate that coupled models, such as BATS–ABL, can be sensitive to the arbitrary choice of the interface height between two model components if they use different assumptions regarding near-surface turbulence. In the case of BATS–ABL, however, this was shown not to compromise that model's ability to investigate the validity of hypothetical aggregation rules.

This study also provides insights into the aggregation of the states (temperature and soil moisture) of an advanced land surface model. Linear aggregation of temperature and soil moisture did not pose any problems as long as vegetation in each patch was not drought-stressed. Aggregation of wet and dry soil patches proved to be a problem because root resistance is a non-linear function of soil moisture in BATS and most probably in the real world. The deterioration of the proposed aggregation scheme in situations with strong soil moisture variability is due to real differences in the calculated, area-average turbulent fluxes. This deterioration is not viewed as a failure of vegetation parameter aggregation *per se*, but rather as a problem associated with variability in near-surface meteorology (if irrigation can be considered a form of precipitation).

It is clear that sharp contrasts in moisture can compromise aggregation, but whether or not they are likely to do so is an entirely separate matter. Variability in soil texture, vegetation, topography, irrigation, and groundwater discharge are all factors that contribute to soil moisture variability. Irrigation was the mechanism that produced variability in this particular study, but irrigated land constitutes significantly less than 1% of the land cover in the contiguous United States, according to the US Geological Survey/EROS Data Center (USGS/EDC) 1-km cover classification of Loveland et al. (1991). In some places, the fraction of irrigated crop is locally higher, though it does not exceed about 10% of the grid area for commonly used GCMs. Model simulations showed that for fractional irrigation cover of 10% or less, the impact of the failure in aggregation rules described here is negligible.

In this study, the extreme differences in soil moisture required to cause failure of aggregation methods did not arise as a result of different vegetation cover (providing

land surface models. The simple aggregation rules, were successful at producing a variety of realistic hydrological (FIFE site). By successful, we mean that are reasonably similar to patches. This success is an effect of subgrid scale, near-surface may be insignificant in the model-related parameters. The vegetation parameter aggregation for adoption of the simple aggregation rules, the results in this study. Therefore, the initiation data from different parameters.

that coupled models, such as the interface height between the surface and near-surface processes shown not to compromise the aggregation rules.

of the states (temperature and aggregation of temperature vegetation in each patch was proved to be a problem in moisture in BATS and most coupled aggregation scheme in to real differences in the is not viewed as a failure is a problem associated with can be considered a form of

compromise aggregation, but a separate matter. Variability in groundwater discharge are all the same was the mechanism that the land constitutes significant land, according to the United States, according to the (EDC) 1-km cover classification of irrigated crop is locally in an area for commonly used vegetation cover of 10% or less, there is negligible.

required to cause failure of vegetation cover (providing

the surface water balance of these cover classes was calculated conservatively); acceptable behavior was always observed for mixes of short and tall grass and agricultural crops. We suspect that this is likely a more general result, especially when seeking aggregate descriptions of the restricted mixture of land covers which are plausible for a particular gridpoint. Equally, it is our suspicion that in most cases the impact of topography and groundwater recharge in creating anomalous areas of moist soil in an otherwise dry landscape is such that the fractional area of these is less than (say) 10%, and that the resulting impact on assumed aggregation rules is therefore limited. The mechanism most likely to generate extended regions of anomalous soil moisture in an otherwise dry landscape is convective rainfall. Convective storms can generate extremely variable rainfall in semiarid environments, and can produce wet patches of soil adjacent to dry patches of soil. Further investigation is required to determine the extent and frequency of such conditions in the real world and whether their presence is significant in terms of spoiling an otherwise simple calculation of the surface energy balance in the semiarid case.

Acknowledgements

It is our pleasure to acknowledge Eleanor Blyth for providing us with the ABL model and helpful advice on its use, and Robert Dickinson for similar provision and advice in the case of BATS. The assistance of Javed Shaikh in coupling these two models was critical. We are thankful to Soroosh Sorooshian for his encouragement and help. The very constructive comments of two anonymous reviewers are greatly appreciated. We are thankful to Corrie Thies for her editorial assistance. We are happy to acknowledge the many scientists who acquired and distributed the FIFE data set, without which this study would not have been possible. The research described in this paper was carried out under NASA Grant number NAGW-3368#1. Altaf Arain was also partially supported by Ministry of Science and Technology, Pakistan.

Appendix A: BATS model description of transpiration under drought stress

If transpiration E_{tr} exceeds the maximum transpiration E_{tr}^{max} that the vegetation can sustain, then stomatal resistance r_s is determined in such a way that $E_{tr} = E_{tr}^{max}$. The maximum transpiration E_{tr}^{max} is calculated by:

$$E_{tr}^{max} = \gamma_0 \sum_i R_{di}(1 - W_{LT}^i) \quad (A1)$$

where γ_0 is the maximum transpiration that can be sustained, with the sum over soil layers each denoted by i ; R_{di} is fraction of roots in a given soil layer; and W_{LT}^i is soil dryness, or plant wilting factor. W_{LT}^i is zero at saturation and unity at permanent

wilting point and is calculated by:

$$W_{LT}^i = \frac{(s_i)^{-B} - 1}{(s_w)^{-B} - 1} \quad (A2)$$

where s_i is the ratio of soil water to saturation in corresponding soil layer, and s_w is the ratio of soil water at the permanent wilting point (i.e. when soil suction reaches 15 bars). B is the factor which defines the change in soil water potential and hydraulic conductivity with soil water content.

Appendix B: Height dependence of drag coefficient

In BATS, a drag coefficient, C_D , is used to calculate sensible and latent heat fluxes. This is a function of the bulk Richardson number R_{iB} and the drag coefficient for neutral stability, C_{DN} which is based on the Monin-Obukhov similarity theory. The bulk Richardson number, R_{iB} , is calculated as:

$$R_{iB} = [gZ_1(1 - T_{surf}/T_a)]/V_a^2 \quad (B1)$$

where $Z_1 = w_i(z_1 - d) + (1 - w_i)z_1$, d is the displacement height of the vegetation, w_i is the fractional vegetation cover, V_a is the wind velocity, T_a is the air temperature, and g is the gravitational acceleration. T_{surf} is the composite surface temperature of vegetation and soil. Over vegetated grid squares, the neutral drag coefficient is estimated as a linear combination of the drag coefficients for the vegetated area C_{DNV} and for the bare soil C_{DNS} (or over snow, if present), thus:

$$C_{DN} = w_i C_{DNV} + (1 - w_i) C_{DNS} \quad (B2)$$

where

$$C_{DNV} = [k/\ln(\{z_1 - d\}/z_{0V})]^2 \quad (B3)$$

and

$$C_{DNS} = [k/\ln(z_1/z_{0S})]^2 \quad (B4)$$

where z_1 is the interface height between BATS and the atmospheric model, k is van Karman constant, and z_{0V} and z_{0S} are the roughness lengths of the vegetation and soil, respectively. The drag coefficient between canopy level and height z_1 is calculated as:

$$\begin{aligned} C_D &= C_{DN}(1 + 24.5(-C_{DN}R_{iB})^{0.5}) \quad \text{for } R_{iB} < 0 \\ C_D &= C_{DN}(1 + 11.5R_{iB})^{-1} \quad \text{for } R_{iB} > 0 \end{aligned} \quad (B5)$$

(A2)

adding soil layer, and s_w is the when soil suction reaches 15 water potential and hydraulic

sible and latent heat fluxes. and the drag coefficient for kkhov similarity theory. The

(B1)

t height of the vegetation, w_i is the air temperature, T_a is the air temperature, T_s is the surface temperature of neutral drag coefficient is for the vegetated area (nt)), thus:

(B2)

(B3)

(B4)

atmospheric model, k is van lengths of the vegetation and vel and height z_1 is calculated

(B5)

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mid elements that range in scale set of vertical calculation such as temperature and describe the soil and

parameters are usually type (called 'cover' here) and that failure to address and may be a significant proposed for addressing parameters, and second. example, Famiglietti and

tion of the land surface parameters are estimated of area-average surface (Shuttleworth, 1991). It is parameters' will be derived the relative proportion of estimates of the parameter ranging procedure should parameter influence.

g fluxes computed with a d with a model describing tions, the performance of d data (e.g. Noilhan and HY, André et al., 1990; , 1994) remain an under- heterogeneity at length ood and Mason, 1991: , which used fairly simple on approach when land ations that do not pene- dary layer (Shuttleworth. where perturbations can provided support for the me et al., 1994; Noilhan

aggregation approach and vegetation characteristics. here model (the Biosphere e and examine aggregation eity. It is a refinement of it uses an advanced land opy of this investigation

was to test various aggregation strategies under realistic conditions. The following guidelines were intended to produce maximum realism in the simulations:

- (a) Select components for the coupled land-atmosphere model which are accepted and realistic: Mason's two-dimensional model of the atmospheric boundary layer (hereafter called ABL model, Mason and Sykes, 1980) and BATS (Dickinson et al., 1986);
- (b) Conduct the analysis using data from a previous large-scale field experiment at times corresponding to periods of intensive study when observed data are available for model initiation and validation: The FIFE experimental site in Kansas, which represents mid-latitude prairie, was selected;
- (c) Investigate the aggregation rules only for vegetation covers defined for BATS and for covers that might realistically be expected to occur in the general vicinity of the experimental site; and
- (d) Examine a variety of field conditions which are considered representative of those observed during the relevant field experiment.

The coupled BATS-ABL model was used to calculate the difference between area-average and aggregate land surface parameterizations under a range of simulated surface covers but using observed initial atmospheric profiles. The BATS-ABL model is described in Section 2; the use of FIFE data is described in Section 3; and various model sensitivity studies are included in Section 4. The results of the aggregation studies are given in Section 5 and discussed in Section 6.

2. The BATS-ABL coupled model

2.1. Atmospheric boundary layer component

The ABL model creates two-dimensional fields of wind velocity, pressure, temperature, and humidity, which are in equilibrium with a heterogeneous land surface once the model has been integrated to steady state. It is a first-order closure model, and the turbulent mixing is assumed to be the same for momentum, heat, and water vapor. The equations used are the two-dimensional form of the ensemble-averaged, Boussinesq approximation to the Navier-Stokes equations, the continuity equation for an incompressible atmosphere, and the conservation equations for humidity and heat. A mixing length formulation is used to derive the Reynolds stress. In neutral and unstable boundary layers, which is the normal situation in this study, the mixing length increases with height up to a maximum of one-third of the boundary layer depth (0.33 km).

The lower boundary conditions of the ABL model are a no-slip condition for wind and values of sensible heat flux and surface humidity which are determined by the land surface model. The upper boundary conditions are no flux of heat and moisture at the top of the boundary layer, i.e. no mixing in the vertical direction at this height. This somewhat artificial condition has little consequence, because the model is usually run forward in time for short periods to allow the lower boundary layer to come into

er values associated with being appropriate for the 2), Stewart and Verma Colello (Personal communication) are given in Table 2.

bserved atmospheric and s. The quality and availability of days for which model experiments selected to were selected in order to litions, and the phase of m the FIFE Information d Sugita (1992), while the ors at several FIFE sites Ball (1993) subsequently n in space (to give a site

nde profiles of potential and pressure, supplement- The initial profiles were alize nine state variables e 3). Because the BATS's fluxes are very sensitive re obtained by running a variables at the dates and

Table 2

BATS soil parameters used in this study

(a) Soil texture parameters for class 7	
Parameter	Value
Porosity	0.51
Minimum soil suction (mm)	200
Saturated hydraulic conductivity (mm s^{-1})	0.0045
Ratio of saturated thermal conductivity of that to loam	0.95
Exponent B	6.8
Moisture content relative to saturation at which transpiration ceases	0.378
(b) Albedo for soil color class 5	
Condition	Albedo
Dry soil, wavelength $< 0.7 \mu\text{m}$	0.16
Dry soil, wavelength $> 0.7 \mu\text{m}$	0.32
Wet soil, wavelength $< 0.7 \mu\text{m}$	0.08
Wet soil, wavelength $> 0.7 \mu\text{m}$	0.16

times of interest. Four such stand-alone simulations were performed to obtain the initial state variables relevant to each cover type used in the study. The stand-alone model was run from May 26, 1987, to October 11, 1987, using observed forcing variables at 30-min intervals (see Table 3 for forcing variables).

4. Model tests

A variety of tests were performed prior to (and during) the aggregation simulations to determine if the model was operating correctly. These are over viewed in the following sections.

4.1. Accuracy of calculated fluxes

Simulated fluxes from the stand-alone version of BATS were compared to observed fluxes obtained from Betts and Ball (1993). Both the observed forcing variables and the observed fluxes are site averages. Tall grass cover type, which corresponds most closely to the actual vegetation at the FIFE site, was used for this comparison. The stand-alone BATS simulation ran for 143 days in the summer of 1987, and calculated fluxes were compared to observed fluxes for the 57 days for which flux measurements are available. The mean difference between simulated and observed fluxes was 6 W m^{-2} for latent heat (with simulated evaporation too small), 6 W m^{-2} for sensible heat (model heating of the air too small), and 15 W m^{-2} for ground heat flux (model heating of the ground was too great). At a 30-min time step, the correlation coefficient between simulated and observed fluxes was 0.80 for latent heat, 0.80 for sensible heat,

Class	Long grass	Irrigated crop
	0.8	0.8
	0.3	0.6
	0.1	0.06
	1	1
	0.1	0.1
	0.8	0.3
	0.08	0.08
	0.3	0.28
200	200	200
6	6	6
0.5	0.5	0.5
2	2	2
5	5	5
0.02	0.02	0.02
0	0	0

4.3. Interface height sensitivity tests

As mentioned in Section 2, the calculation of surface energy fluxes may be sensitive to the arbitrary selection of the interface height between the two submodels in BATS-ABL. To evaluate this possibility, integrations were carried out with the BATS-ABL model using two very different interface heights, namely 2 m (the standard height used in this study) and 20 m (the standard height of tall vegetation in BATS). Fig. 2 illustrates the initial observed profiles over short grass for noon on August 15, 1987, and the ensuing atmospheric profiles calculated by BATS-ABL after running for 12 s and 10 min. BATS-ABL smooths the profile and 'retunes' simulated atmospheric profiles so that they are in equilibrium with the surface. It appears that BATS provides a more rapid analytic integration of the variations between 2 and 20 m than the finite difference integration used in ABL and, in consequence, the

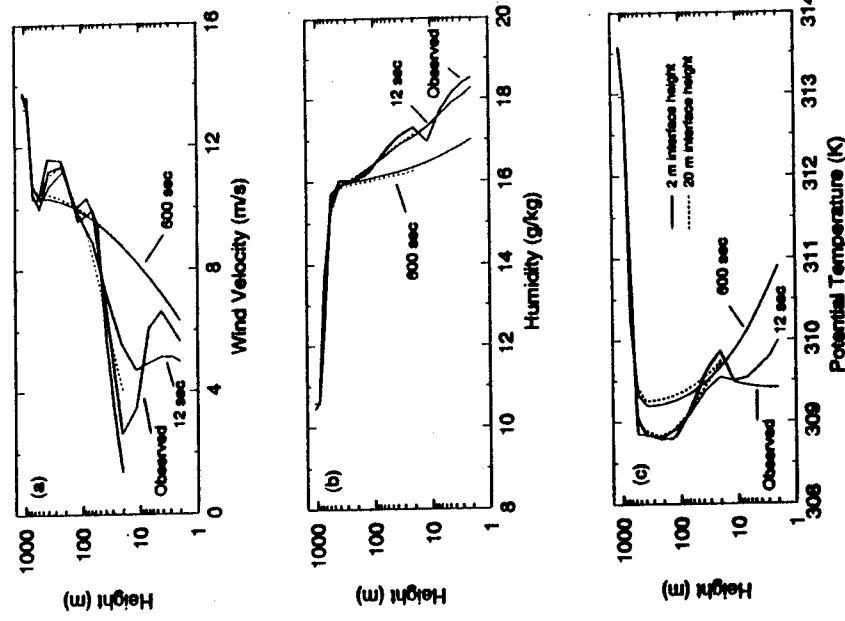


Fig. 2. Profiles of (a) wind, (b) humidity, and (c) temperatures at noon on August 15, 1987. Initial observed profiles are shown, along with simulated profiles for short grass. Simulated profiles were generated by BATS-ABL after running for 12 s and 600 s with the interface height set to 2 m and 20 m.

While the observed and
es were sufficiently similar
TS as a reasonable descrip-

ed initial conditions, there
eric model are adjusting to
ned to determine length of
s. Unless otherwise stated,
of 600 s.

dependency on meteorological variables does not seem to compromise the effectiveness of this cancellation significantly.

5. Definition and tests of aggregation rules

5.1. Aggregation rules for BATS parameters

The purpose of an aggregation rule is to define an "effective parameter which is the weighted average of component covers, through that function involving the parameter which most nearly links it to a linear average of the associated flux" (Shuttleworth, 1991). Averaging rules for most parameters are thus based on the manner in which they influence fluxes. For example, because solar radiation is related linearly to albedo, it follows that albedo should be aggregated as an area-weighted arithmetic average. The aggregation rules used in this study are defined in Table 4. Area-weighted arithmetic averaging was performed for all parameters except aerodynamic roughness length, z_0 , and stomatal resistance, r_s .

Roughness length and stomatal resistance are of primary importance in determining surface fluxes. Roughness is averaged in such a way that the effective value equals that value for which a homogeneous area would provide a surface stress equal to the spatial average of a heterogeneous area. The averaging is based on the blending height (Wierenga, 1986; Mason, 1988). The value of blending height (in m) can be calculated by the equation given in Table 4 or approximated by the patch length (in m) divided by 200 (Mason, 1988).

Leaf stomata are the primary route for the flow of water vapor from the near-saturated surface conditions within the leaf to the canopy air outside. In BATS, a minimum stomatal resistance, $r_{s \min}$, is attenuated using multiplicative stress functions

Table 4

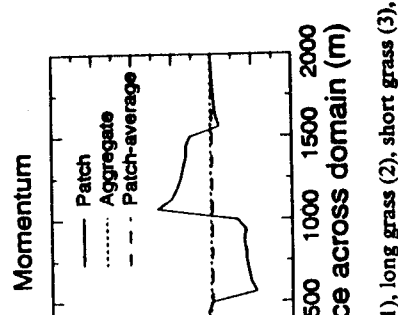
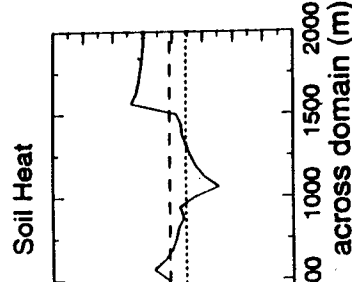
Aggregation rules for BATS vegetation parameters

Parameter	Aggregation rule	Comment
Maximum leaf area index (LAI _{MAX})	$LAI_{MAX} = \sum w_i LAI_{MAX_i}$	Linear effect on radiation and evaporation
Minimum stomatal resistance ($r_{s \min}$)	$\frac{1}{r_{s \min}} = \sum \frac{w_i}{r_{s \min i}}$	(i) Reciprocal rule: evaporation inversely related to stomatal resistance in BATS
	$r_{s \min} = \sum w_i r_{s \min i}$	(ii) Linear rule
Roughness length for momentum (z_0)	$\ln \left(\frac{z_b}{z_0} \right)^2 = \sum w_i \ln \left(\frac{z_{b_i}}{z_0} \right)^2$	Blending height given as $z_b = 2(u^*/V_s)^{-1} l_c$ where l_c is the horizontal length scale

The subscript "i" represents cover (vegetation) type and "w_i" represents the fractional area of that cover type. Parameters (see Table 3 for a list) that use a linear area-weighted average have been omitted for brevity, with the exception of maximum leaf area index, which is included as an example.

alternate 1-km patches, for equal time periods little difference in calculations in a 2-km domain were seen in tests.

This four-patch simulation increased evaporation at the cool air from Patch 1 (the edge of Patch 2 is balanced by evaporation at the left edge of Patch 1). BATS has an assumed



(1), long grass (2), short grass (3).

and root soil moisture were were 15.5 and 314.1 mm, and aggregate vegetation which an artificially wet unrealistically assumes that specifically, with the noon when irrigated crops were crop.

agation rules worked well moisture. This would be d sensible heat are small, even when there were quite between the two patches. ot significant because the or each of the simulations.

irrigated crop transpired at mixed crop and long grass) vector, W_{LT} , plotted against of the soil moisture ratio, evaluated for a wet soil is ture average. The effect of and irrigated crop for the oil moisture for the aggre- patches. The latent heat r than the average of latent

two crops

a cover



ressed as a fraction of saturation. October 11, 1987.

Table 5

Soil moisture and turbulent fluxes for October 11, 1987 (noon) simulation

	Mixed crop	Irrigated crop	Aggregate cover
Upper soil moisture	18 mm	36 mm	27 mm
Root zone soil moisture	220 mm	371 mm	295 mm
Latent heat flux	95 W m ⁻²	276 W m ⁻²	262 W m ⁻²
Sensible heat flux	245 W m ⁻²	131 W m ⁻²	(Patch average = 185) 173 W m ⁻² (Patch average = 188)

The failure of aggregation in these simulations, which is caused by a non-linearity in root resistance, results in a real difference in the net turbulent fluxes returned to the atmosphere (374 W m⁻² for the average of the two patches versus 435 W m⁻² for the aggregate simulation). It should be noted that such a result would not be obtained for models which assume zero ground heat flux. In the case of the BATS model, it would be possible to avoid this aggregation problem by calculating the reciprocal average of root resistance, and then determining the aggregate soil moisture corresponding to that aggregate root resistance; however this is not an acceptable solution because doing so would break the water balance. There is a clear need to determine whether, in the real world, spatial differences in soil water availability can and do occur with sufficient regularity and with sufficient extent for this to be a significant issue in meteorological models. We return to this point in Section 6.

5.4. Alternate aggregation rules for stomatal resistance

In this study, several aggregation rules were examined for stomatal resistance, because previous studies have found support for both linear and reciprocal averaging rules, while Blyth et al. (1993) found that an average of linear and reciprocal averages gave the best results. However, in practice, there is little difference in the calculated surface energy balance irrespective of the selected aggregation rule. For noon on August 15, for example, there was a 2% difference between the latent heat obtained with the linear average and the latent heat obtained from the reciprocal average, despite the fact that the latent heat flux for individual patches differs by about 100 W m⁻² (see Fig. 4). This difference was 1% for comparisons involving mixed crop and irrigated crop for noon on October 11. The results are similar for other fluxes and at other times. Therefore it appears that either the linear average or the reciprocal average (or the average of these two) works equally well in providing an estimate of the aggregate surface resistance in the case of the BATS land covers which are plausible in Kansas.

6. Summary and discussion

This paper provides additional support for the use of simple aggregation rules

and surface models. The simple aggregation rules, however, were successful at producing a variety of realistic hydrological (FE site). By successfully aggregating what are reasonably similar patches, this success is a function of subgrid scale, near-surface, and is insignificant in the context of non-related parameters. The aggregation parameter aggregation for adoption of the simple aggregation rules, the aggregation in this study. Therefore, the aggregation data from different parameters.

At coupled models, such as the interface height between the surface and near-surface, shown not to compromise the aggregation rules. The states (temperature and aggregation of temperature) in each patch was proved to be a problem. Moisture in BATS and most aggregation schemes in the real differences in the aggregation is not viewed as a failure. A problem associated with aggregation is considered a form of compromise aggregation, but an aggregation matter. Variability in groundwater discharge are all on was the mechanism that land constitutes significant States, according to DC) 1-km cover classification of irrigated crop is locally used area for commonly used aggregation cover of 10% or less, there is negligible. The aggregation required to cause failure of vegetation cover (providing

the surface water balance of these cover classes was calculated conservatively); acceptable behavior was always observed for mixes of short and tall grass and agricultural crops. We suspect that this is likely a more general result, especially when seeking aggregate descriptions of the restricted mixture of land covers which are plausible for a particular gridpoint. Equally, it is our suspicion that in most cases the impact of topography and groundwater recharge in creating anomalous areas of moist soil in an otherwise dry landscape is such that the fractional area of these is less than (say) 10%, and that the resulting impact on assumed aggregation rules is therefore limited. The mechanism most likely to generate extended regions of anomalous soil moisture in an otherwise dry landscape is convective rainfall. Convective storms can generate extremely variable rainfall in semiarid environments, and can produce wet patches of soil adjacent to dry patches of soil. Further investigation is required to determine the extent and frequency of such conditions in the real world and whether their presence is significant in terms of spoiling an otherwise simple calculation of the surface energy balance in the semiarid case.

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It is our pleasure to acknowledge Eleanor Blyth for providing us with the ABL model and helpful advice on its use, and Robert Dickinson for similar provision and advice in the case of BATS. The assistance of Javed Shaikh in coupling these two models was critical. We are thankful to Soroosh Sorooshian for his encouragement and help. The very constructive comments of two anonymous reviewers are greatly appreciated. We are thankful to Corrie Thies for her editorial assistance. We are happy to acknowledge the many scientists who acquired and distributed the FIFE data set, without which this study would not have been possible. The research described in this paper was carried out under NASA Grant number NAGW-3368#1. Altaf Arain was also partially supported by Ministry of Science and Technology, Pakistan.

Appendix A: BATS model description of transpiration under drought stress

If transpiration E_{tr} exceeds the maximum transpiration $E_{tr}^{\max}$ that the vegetation can sustain, then stomatal resistance r_s is determined in such a way that $E_{tr} = E_{tr}^{\max}$. The maximum transpiration $E_{tr}^{\max}$ is calculated by:

$$E_{tr}^{\max} = \gamma_{r0} \sum_i R_{di} (1 - W_{LT}^i) \quad (A1)$$

where γ_{r0} is the maximum transpiration that can be sustained, with the sum over soil layers each denoted by i ; R_{di} is fraction of roots in a given soil layer; and W_{LT}^i is soil dryness, or plant wilting factor. W_{LT}^i is zero at saturation and unity at permanent

(A2)

inding soil layer, and s_w is the
hen soil suction reaches 15
er potential and hydraulic

sible and latent heat fluxes.
nd the drag coefficient for
khov similarity theory. The

(B1)

height of the vegetation, w_i
, T_a is the air temperature,
site surface temperature of
neutral drag coefficient is
nents for the vegetated area
t), thus:

(B2)

(B3)

(B4)

atmospheric model, k is van
lengths of the vegetation and
el and height z_1 is calculated

(B5)

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MAPPING SURFACE COVER PARAMETERS USING AGGREGATION RULES AND REMOTELY SENSED COVER CLASSES

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Abstract

A coupled model, which combines the Biosphere-Atmosphere Transfer Scheme (BATS) with an advanced atmospheric boundary-layer model, was used to validate hypothetical aggregation rules for BATS-specific surface cover parameters. The model was initialized and tested with observations from the Anglo-Brazilian Amazonian Climate Observational Study and used to simulate surface fluxes for rain forest and pasture mixes at a site near Manaus in Brazil. The aggregation rules are shown to estimate parameters which give area-average surface fluxes similar to those calculated with explicit representation of forest and pasture patches for a range of meteorological and surface conditions relevant to this site, but the agreement deteriorates somewhat when there are large patch-to-patch differences in soil moisture. The aggregation rules, validated as above, were then applied to remotely sensed 1 km land cover data set to obtain grid-average values of BATS vegetation parameters for $2.8^\circ \times 2.8^\circ$ and $1^\circ \times 1^\circ$ grids within the conterminous United States. There are significant differences in key vegetation parameters (aerodynamic roughness length, albedo, leaf area index, and stomatal resistance) when aggregate parameters are compared to parameters for the single, dominant cover within the grid. However, the surface energy fluxes calculated by stand-alone BATS with the 2-year forcing data from the International Satellite Land Surface Climatology Project (ISLSCP) CDROM were reasonably similar using aggregate-vegetation parameters and dominant-cover parameters, but there were some significant differences, particularly in the western USA.

1. Introduction

Natural surfaces vary from place to place with a scale of heterogeneity which ranges from meters to hundreds of kilometers. Parameterizing such heterogeneous land surfaces in weather and climate prediction models is problematic because the grid square typically represents an area between 10^4 and 10^6 km^2 . In recent years, efforts have been made to improve the description of heterogeneous land surfaces in these models and there has been progress towards defining averaging procedures to calculate area-average land surface exchanges at local and regional scales (Mason, 1988; Wood and Mason, 1991; Blyth et al., 1993; Lhomme et al., 1994; Lynn et al., 1995). However, only a few past studies (such as Noilhan and Lacarrère, 1995) have validated model performance against observations, and little research has used the land surface schemes actually in use within General Circulation Models (GCMs).

One way to describe heterogeneous vegetation is to use a standard land-surface scheme with effective, area-average vegetation parameters calculated using simple aggregation rules applied to all the vegetation cover types present in a grid element. Aggregation rules (Shuttleworth, 1991) are simple averaging procedures to calculate effective values of surface cover parameters which estimate area-average surface fluxes similar to those given with explicit representation of the separate patches of vegetation. A weighted linear average of surface-cover parameters based on the fractional area of all the cover types present in the grid is the simplest averaging procedure, but if fluxes are not proportional to parameters (such as the aerodynamic roughness length), then linear averaging is inappropriate.

The Biosphere-Atmosphere Transfer Scheme (BATS, Dickinson et al., 1986) is the land-surface parameterization used in the National Center for Atmospheric Research's (NCAR's) Community Climate Model. In normal application of BATS, the values of vegetation parameters are based on the single dominant surface-cover in each grid cell. A set of hypothetical aggregation rules for BATS surface cover parameters are given in Table 1 (Arain et al., 1996). Linear averaging is proposed for all parameters except aerodynamic roughness length and minimum stomatal resistance. The aggregate value of aerodynamic roughness length is calculated by averaging the drag coefficient from Mason (1988). Not shown in Table 1 are the three alternative averaging procedures that have been proposed for minimum stomatal resistance: linear averaging, reciprocal averaging, and average of these two averages.

Arain et al. (1996) coupled BATS with an advanced, two-dimensional model of the atmospheric boundary-layer first developed by Mason and Sykes (1980) to obtain a realistic coupled model of surface-atmosphere exchanges. This model, hereafter called the BATS-ABL model, was used to investigate the acceptability and applicability of aggregation rules for BATS vegetation parameters using observed forcing and surface data from the First International Satellite Land Surface Climatology Project Field Experiment (FIFE) site in Kansas (Sellers et al., 1992). The results showed that the aggregation rules successfully calculate area-average surface fluxes under a variety of hydrometeorological conditions at this site. However, they did not work well when artificially wet (irrigated) patches were combined with naturally dry soil patches.

In the Arain et al. (1996) study, the application of aggregation rules for some important BATS parameters such as aerodynamic roughness length, zero plane displacement height, and minimum stomatal resistance was poorly tested because there is little difference in these parameters for the cover classes relevant to the FIFE site. Therefore it is appropriate to extend aggregation tests to those mixes of vegetation which have a large contrast in these three parameters. Cleared areas within undisturbed Amazonian forest provide the desired contrast. The results of aggregation tests carried out for adjacent patches of Amazonian evergreen broadleaf trees and short grass are presented in Section 2 of this report. Observed data used for model initiation, forcing, and validation were obtained from the Anglo-Brazilian Climate Observational Study (ABRACOS) (Shuttleworth et al., 1991; Gash et al., 1996).

On the basis of the success reported in the current (and previous) study, the BATS-specific aggregation rules were then applied to the 1 km land-cover data set of Loveland et al. (1991) to calculate grid-average values of BATS vegetation parameters for $2.8^\circ \times 2.8^\circ$ and $1^\circ \times 1^\circ$ grid squares across the conterminous United States (U.S.). A comparison of the ensuing aggregate parameters with parameter values given by the conventional dominant cover within a grid cell are described in Section 3. In Section 4, the forcing data from the International Satellite Land Surface Climatology Project (ISLSCP) "Initiative 1" (Meeson et al., 1995) are used to conduct 2-year BATS simulations for the conterminous U.S. at a 1° resolution. Fluxes from aggregate parameters are compared to fluxes from parameters based on the dominant vegetation type in each grid.

2. Testing Aggregation Rules in the Amazon

2.1 The ABRACOS Experiment

The Anglo-Brazilian Climate Observational Study (ABRACOS) sought to improve the parameterization of surface-atmosphere models by providing accurate and representative data for forested and deforested areas in the Amazon. Detailed studies of surface climatology, micrometeorology, plant physiology, and soil hydrology were made at three different forest and adjacent clearing sites across the Amazon (Shuttleworth et al., 1991; Gash et al., 1996).

Data from the *Reserva Duke* forest site and *Fazenda Dimona* pasture site, which are located respectively about 25 and 100 km from Manaus, Amazonas, were used in the current study. *Reserva Duke* (2° 57'S, 59° 57'W) is a protected area, and the experimental site is surrounded by undisturbed forest for at least 5 km in all directions. The mean canopy height is 35 m, but some trees reach 40 m, and meteorological data were collected at the top of a 45m tower. Detailed descriptions of the site and measurements have been given by Shuttleworth et al. (1984), Roberts et al. (1990), Shuttleworth et al. (1991) and Gash et al. (1996). *Fazenda Dimona* (2° 19'S, 60° 19'W) is a 10 km² forest clearing which was created by felling and burning the original forest and then planting pasture grasses such as *Brachiaria decumbens* and *Brachiaria humidicola*. About 11% of the total area is bare soil, while 5% area is covered with tree trunks (Wright et al., 1992).

For logistic reasons, soil moisture data were recorded at a forest site near the *Fazenda Dimona* pasture site rather than at the *Reserva Duke* site. The soil at *Fazenda Dimona* is a yellow Latosol (*Oxisol* or *Haplic Acrorthox*) with 80% fine clay content and high conductivity (Wright et al., 1995). The soil moisture data were collected weekly at 20cm intervals to a depth of 2 m using neutron probes. Hourly measurements of incident solar radiation, wind speed, air temperature, specific humidity, and precipitation were made with automatic weather stations. Net radiation, sensible heat, latent heat, and soil heat fluxes were also measured at the pasture site, while only net radiation and soil heat flux were recorded at the forest site. At both sites, data collection started in mid-September 1990, and continued until mid-September 1991, although there were times when some of the meteorological and soil moisture data were missing at one or both sites during this period.

2.2 Validation of BATS at the Amazon Sites

Short grass and evergreen broadleaf trees are the appropriate BATS surface cover classes for the pasture and forest site, respectively. The corresponding set of morphological, physical and biophysical parameters are given in Table 2. Based on ABRACOS observations (Wright et al., 1996), soil textural class 10 and soil color class 3 were selected as suitable for the Manaus site. The associated soil parameter values are given in Table 3.

A stand-alone version of BATS was run using the observed hourly values of incident solar radiation, air temperature, specific humidity, wind velocity, and precipitation as forcing data. It is

necessary to initiate nine state variables (see Table 4). Surface soil temperature, deep soil temperature, canopy temperature, and canopy air temperature were all initialized to observed air temperature in BATS. The ground wetness factor was set to 1, while the volumetric soil moisture in the three BATS soil layers was obtained by linear interpolation of the observed data. The soil moisture values observed at 2 m were assumed to be the same as at 10 m, which is the depth of the lowest BATS soil layer.

Surface flux data for the pasture site were available for validation for 27 days in 1990 and 71 days in 1991. When the above-described default parameters were used in the stand-alone version of BATS, the calculated latent heat fluxes were, on the average, 38 Wm^{-2} too low, while sensible heat fluxes were too high by 18 Wm^{-2} . The root mean squared error between 30-minute model estimates and observations of these fluxes were 82 Wm^{-2} and 57 Wm^{-2} , respectively (see Table 5a). In this study, the philosophy adopted was to make the minimum parameter changes necessary to achieve acceptable agreement between observations and model calculations. In practice, the simulations of the pasture energy fluxes were greatly improved merely by changing the value of minimum stomatal resistance from 200 sm^{-1} to 40 sm^{-1} . With this change, the mean bias was reduced to 11 Wm^{-2} for latent heat and 4 Wm^{-2} for sensible heat, while the corresponding root mean squared errors reduced to 54 and 40 Wm^{-2} , respectively (see Table 5b). These improvements for July 21, 1991, a day on which the soil moisture was fairly high are illustrated in Figure 1a-b.

Forest validation runs were also performed over the two time intervals 3 October 1990 to 31 January 1991 and 1 July 1991 to 26 September 1991, these being the times when both soil

moisture and meteorological forcing data were available at the *Reserva Ducke* forest site. Although surface heat fluxes were not available at this site for this period, other studies for forested areas in the Amazon have shown that, on the average, about 80% of the available energy is used for transpiration, and that there is no substantial decrease in evapotranspiration during low rainfall periods, (see, for example, Shuttleworth, 1988 and Wright et al. 1996). BATS simulations using default parameters gave latent heat fluxes which were too small during periods of low rainfall.

The default parameters for evergreen broadleaf forests (given in Table 2) assume the rooting depth for forest cover is 1.5 m and that 80% of these roots are in the upper 10 cm of soil. However, there is strong field evidence (e.g., Hodnett et al., 1995; Nepstad et al., 1994) that forest roots penetrate and extract water to depths of at least 3.5 m in the Amazon. To reflect this, the depth of the rooting zone in BATS was increased from 1.5 m to 4.0 m, and the distribution of roots was changed so that only 20% (rather than 80%) of roots were assigned to the upper 10 cm deep soil layer. This change greatly improved the simulated surface fluxes for the forest in dry conditions. These improvements are shown in Figures 1c-d.

The aggregation studies described in the next section were made using both the original "default" parameters and the revised parameters to establish the sensitivity or otherwise to the assumed parameter values.

2.3. Aggregation Studies

2.3.1 Procedures

The procedures used to test aggregation rules were essentially identical to those described in detail by Arain et al. (1996). Two versions of the BATS-ABL model were compared: one with explicit representation of a 1-km long pasture adjacent to a 1-km long forest, and one in which vegetation is a homogeneous blend of pasture and forest. In the homogeneous case, vegetation parameter values were based on the aggregation rules listed in Table 1 applied to a 50% pasture, 50% forest mix. Both versions of the model were initiated with identical observed meteorological and soil moisture data. The models were run forward for 10 minutes at particular times on selected days. A comparison was then made between model-calculated area-average energy fluxes of the heterogeneous and homogeneous models. The level of agreement between these two different calculations is then used to evaluate the performance of the aggregation rules.

In practice, the simulated fluxes (especially those of soil heat flux) are sensitive to the initial assignment of deep soil temperature because the model is run forward only for a short period of time. Hence, the initial state variables used in the model were obtained by selecting the required values at appropriate times from multi-month stand-alone BATS simulations for pasture, forest, and aggregate-vegetation.

The interface height between BATS and the atmospheric boundary-layer model was set to 45 m. [For a detailed discussion of the sensitivity of simulated fluxes to the choice of this model

interface height, see Arain (1994) and Arain et al. (1996)]. Initial vertical profiles of potential temperature, humidity, and wind speed were generated by running a one-dimensional version of BATS-ABL model forward for several hours with fixed surface conditions until the calculated values of these atmospheric variables matched the observation 45 m above the ground at the forest site. It was assumed that the remainder of calculated vertical profiles were then in equilibrium with area-average surface exchanges.

2.3.2 Results

Two-patch and corresponding aggregate-cover simulations using the BATS-ABL model were made on five cloud-free days three times a day: one mid-morning, one close to noon, and one mid-afternoon (see Table 6). Many aspects of climate are fairly similar throughout the year at this ABRACOS site, but rainfall is seasonal. By chance, one dry period during the 1990 data collection was much longer than is usual, and this gave an excellent opportunity to observe effects of dry periods on aggregate surface behavior with these two very different surface vegetation covers. Two study days were selected from this unusually dry period and three from a more typical wet period--these judgments being based on the observed soil moisture. The mid-day observed soil moisture values in the upper 0.1 m soil layer and in the 1 m rooting layer are given in Table 6.

October 26, 1990 was a hot, sunny day when the soil moisture in the rooting zone was low. Simulated evaporative flux from the forested area was large and showed no evidence of soil moisture stress because of the deeper roots, while evaporation from the pasture patch was low. In

the BATS-ABL model, the corresponding aggregate vegetation cover also did not show evidence of soil moisture stress because of its (assumed) 2.5 m deep rooting depth. The simulated two-patch and aggregate-cover area-average energy fluxes were noticeably different on this day, especially for the noon and mid-afternoon runs. October 30, 1990 was also a dry day with low soil moisture in the rooting zone, and the difference in evaporative flux from forest and pasture patches was again as large as 271 Wm^{-2} at mid-day. In fact, the pasture patch was close to wilting point, while there was still plenty of water available for evaporation in the deeper-rooted forest patch. The variation in sensible heat, latent heat, soil heat, and momentum flux across the modeled domain from two-patch and homogeneous cover is shown in Figure 2. The differences between patch average and aggregate fluxes were significant (up to 25 Wm^{-2}).

December 10, 1990 was a wet day at the beginning of the rainy season in this part of the Amazon. Soil moisture at the pasture site was sufficiently high to essentially eliminate the soil-moisture control on transpiration low, so that the pasture was evaporating at close to potential rate. The two-patch and aggregate-cover area-average energy fluxes were similar for all three runs on this day. The last two days on which BATS-ABL runs were made, July 3 and August 28, 1991, were days with high soil moisture. The evaporative fluxes from both the pasture and forest patches were large, and the latent heat from aggregate cover was also high. The proposed BATS-relevant aggregation rules again worked well at all three times on these two days.

Comparison of patch-average and corresponding aggregate sensible heat, latent heat, soil heat and momentum flux for all 15 BATS-ABL runs is given in Figure 3. In these runs, the revised vegetation parameters were used (see Section 2.2). This figure shows that the aggregation rules

worked quite well when neither of the vegetation patches has soil moisture stress, but agreement between two-patch and aggregate-cover area-average fluxes deteriorates somewhat when one of the patches (the shorter-rooted pasture patch) is subject to soil moisture stress. Two-patch and corresponding aggregate-cover runs were also performed for the above-mentioned days and tithes using the BATS default vegetation parameters. The results, which are shown in Figure 4, are broadly similar.

Hence, the overall results of model experiments at the ABRACOS (Manaus) site are in essence the same as those reported by Arain et al. (1996) for the FIFE site. At the FIFE site, aggregation rules did not work well when dry soil moisture patches were combined with artificially wet (irrigated) patches. They also perform less well at the ABRACOS site when there is a large soil moisture deficit in the shallower-rooted pasture patch, while the forest patch (with deeper roots) still has access to a substantial moisture store.

In order to investigate the sensitivity of aggregation rules to large differences in soil moisture in adjacent patches in greater detail, the stand-alone BATS runs for forest and pasture patches were compared with a run for the corresponding aggregate cover using observed forcing data at the *Reserva Duke* forest site. Figures 5a-b show daily average energy fluxes from two-patch and corresponding aggregate-cover runs for periods with dry and wet soils, while Figure 5c shows the wilting factors (calculated within the BATS model for upper and root soil layers) for both pasture and forest patches for the entire study period. The plant wilting factor is given by:

$$w_{LT}^i = \frac{(s_i)^{-B} - 1}{(s_w)^{-B} - 1}$$

where s_i is the ratio of soil water at saturation for ith soil layer, and s_w is the ratio of soil water at the permanent wilting point (i.e., when suction reaches 15 bars), while B ($= 9.2$) is the factor which is used to define the change in soil water potential and hydraulic conductivity with soil moisture.

As expected, the differences in sensible heat, latent heat, and soil heat flux between heterogeneous and aggregate cover is most pronounced during the time period when there was a large soil moisture deficit (see Figure 5a and 5c). In fact, the upper soil is close to the wilting point for both the pasture and forest during this unusually dry period. In the rooting zone, however, the pasture is drought stressed and the forest is not. Because there is a non-linear relationship between the wilting factor and soil moisture content given in Equation (1), the average root soil moisture does not lead to the average wilting factor. However, it is reassuring to note that, even with this extreme difference in cover type (resulting in a four-fold difference in rooting depth), the discrepancy between daily average surface energy fluxes is still only about 10% or 10 Wm^{-2} in dry soil conditions which are, at least in the case of this particular ABRACOS study site, uncommon and short-lived.

Table 1 gives three alternative aggregation rules for combining minimum stomatal resistance, these being linear averaging, reciprocal averaging and, following the suggestion of Blyth et al. (1993), the average of these last two averages. Evaluation of their relative merits can be made at this ABRACOS site because applying these three rules with the preferred value of minimum

stomatal resistance for short grass (40 m^2) and evergreen forest (150 m^2) gives the values of 95, 63, and 79 m^2 , respectively. Model runs were made using both BATS-ABL and the stand-alone version of BATS with these three alternative values for the aggregate value of minimum stomatal resistance. Typical results for the stand-alone BATS model are illustrated in Figure 6. Figure 6a shows calculated fluxes on October 30, 1990, a day with dry soil, and Figure 6b shows calculated fluxes on August 28, 1991, a day with wet soil. Applying the three alternative rules gave results which (although not good) are very similar to each other when soils are dry; however in more common wet soil conditions, the linear aggregation rule consistently performed best. Similar results were obtained from simulations using coupled the BATS-ABL model on the five days and three times of the day for which tests were made. On the basis of this evidence, a linear aggregation rule is recommended for this BATS parameter, and linear aggregation of minimum stomatal resistance was adopted and applied for the remainder of the analysis described below.

3. Application of Aggregation Rules for USA

Results for the ABRACOS site and those reported by Arain et al. (1996) for the FIFE site suggest that the proposed BATS-specific aggregation rules given in Table 1 calculate area-average energy fluxes that are acceptably similar to those calculated with explicit representation of separate vegetation patches under a wide range of hydrometeorological conditions and for a range of surface covers appropriate to these two sites. Shortcomings can arise, but these are not because of patch-to-patch differences in the vegetation-related parameters *per se*; rather, they occur if there are marked differences in the plant-accessible soil moisture in the heterogeneous

landscape, and they arise because the root resistance assumed in BATS is a non-linear function of soil moisture.

In the case of the study at the FIFE site (Arain et al., 1996), these shortcomings were substantial when artificially wet (irrigated) soil patches were combined in equal proportion with naturally dry soil patches. However, an analysis (similar to that described later) of irrigated land falling within any individual grid square used in meteorological models shows that the proportion is rarely, if ever, greater than 10% in the USA--the USA being the only region for which satellite-derived data on land cover are readily available at this time. Moreover, coupled modeling studies using BATS-ABL show that including artificially irrigated vegetation with this relative extent still allows the use of aggregation rules with acceptable accuracy in the calculated area-average fluxes. Hence, the presence of patches of artificially irrigated soil is not usually a serious issue. In the study at the ABRACOS site described above, this same feature is again revealed (albeit in a less significant way) when patches of deep-rooted forest are combined with patches of shallower-rooted pasture in periods of exceptionally low rainfall.

For the purposes of the remainder of this study, we chose to ignore this complicating, indirect influence of heterogeneous soil moisture acting through the overlying vegetation and proceed to investigate the sensitivity of surface parameters and exchanges under the assumption that the BATS-specific aggregation rules for vegetation parameters given in Table 1 have broader acceptability and applicability. However, the reader is cautioned that the following analysis is questionable in two specific cases, namely for landscapes and climates in which there are either (a) large differences in soil moisture and with wet and dry soils covering approximately equal areas and with soil moisture differences greater than 10-20% of the rooting zone soil moisture store; or (b) approximately equal areas of shallow-rooted and deep-rooted vegetation growing in a climate where the shallower-rooted vegetation is under soil moisture stress.

3.1 The USGS/EDC Land Cover Data set

A prototype land-cover characteristics data base for the conterminous United States is available from the U.S. Geological Survey (USGS) EROS Data Center (Loveland et al., 1991) for use in a variety of global modeling, mapping, monitoring, and analytical studies. These data were derived by classifying the 1990 NOAA-11 Advanced Very High Resolution Radiometer time-series data at 1 km x 1 km resolution, with post-classification refinement based on other Earth science data including topography, climate, soils, and ecoregions. The data base uses the concept of seasonal land cover regions as a framework for presenting the temporal and spatial patterns of vegetation. A set of 28-day maximum NDVI composite images covering the conterminous United States were clustered into 70 spectral-temporal classes and subsequently stratified, refined, and labeled using the ancillary data. As a result, 159 seasonally distinct spectral-temporal land-cover classes were labeled according to their constituent vegetation types and, in the case of the data used in this study, these 159 classes were then translated into the 18 BATS-relevant cover classes. An additional nine classes were added which correspond to mixtures of two classes within a 1 km x 1 km pixel.

3.2 Aggregate Parameters from Remotely Sensed Land Covers

We obtained aggregate BATS parameter values for 2.8° GCM grids by computing the area-average of the 1 km BATS values obtained from the USGS-EROS land cover data base described above. Aggregate values of vegetation parameters (Table 3) were computed using the

aggregation rules given in Table 1 (but assuming a linear aggregation rule for minimum stomatal resistance). In this section, the resulting parameter values are compared with the equivalent values for the single most common BATS vegetation cover present in the grid square, this being the method used for assigning parameters in GCMs at the present time. We used a grid size of 2.8° latitude by 2.8° longitude because this is the standard horizontal resolution of NCAR's Community Climate Model version 2 (CCM2) and is a typical resolution of other climate models.

The fractional cover area of all vegetation types within a given grid cell was calculated using the total number of (1 km x 1 km) pixels for each cover class. In the case of cover types assigned a number 20-28 in the USGS-EROS data, 50% of the area of each pixel was assigned to be one cover type and 50% to the other cover type in these calculations. The number of remotely sensed cover classes covering more than 5% in each 2.8° x 2.8° grid square of CCM2 is given in Figure 7a. There can be up to eight such classes in an individual grid square, but more typically there are 3-5 different BATS cover classes present. A histogram of the fractional area covered by the most common cover present in each CCM2 grid across conterminous U.S. is displayed in Figure 7b. On the average, about 50% of the grid square is covered with the most common cover class, but rarely is the grid square entirely covered with one vegetation class. In some cases, the most common vegetation class covers only 20-30% of the grid.

The percentage difference between the value of important BATS-specific vegetation parameters for the single cover currently assigned to each grid square in CCM2 and the value calculated for aggregate-cover using USGS-EROS data is illustrated in Figure 8. Maximum fractional

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vegetation cover is influenced little by using these different approaches of parameterization for most of the USA (see Figure 8a), but there are pronounced differences in some regions of the west. This is because the default cover assigned within CCM2 is semi-desert in these regions, but the remotely sensed data show the substantial presence of vegetation with higher vegetation cover fraction. In contrast, the values assigned to minimum stomatal resistance (Figure 8b) show significant differences in central and eastern portions of the USA, with the sign depending on whether more or less crop land (which is assumed to have lower stomatal resistance than other cover types in BATS) is included in the area-average calculation.

The calculated values of area-average albedo and warm season leaf area index (Figures 8c-d) also show marked differences, especially in the western United States. The use of default cover class assigned in CCM2, which is often short grass or shrub land, neglects the presence of other surface cover types such as desert and evergreen forest. However, the presence of these covers is reflected in the parameter values calculated for aggregate cover. The assigned values of cold season leaf area index and aerodynamic roughness length exhibit marked differences (of up to $\pm 100\%$) across the entire United States (Figures 8e-f), but the spatial distribution of differences in these parameters is difficult to interpret.

Figure 9 is similar to Figure 8 except that aggregate parameters were computed for the $1^\circ \times 1^\circ$ grid mesh. In this case, the most common cover within each grid square is based on the USGS-EROS data set rather than on the cover type currently used in the CCM2 parameterization. In general terms, the percentage differences in parameter values follow a similar pattern to that calculated for the $2.8^\circ \times 2.8^\circ$ (CCM2) grid mesh, but it is easier to interrelate spatial patterns

between parameters in the case of this smaller $1^\circ \times 1^\circ$ grid mesh. The most noticeable distinction between Figure 8 and Figure 9 occurs in the western United States for vegetation fraction, leaf area index, albedo, and minimum surface resistance. This distinction is not related to the use of aggregate covers, but is due to a mismatch between the single cover currently assigned within CCM2 and that measured from satellite.

4. Simulations Using ISLSCP CDROM Data

4.1 Model Forcing Variables and Initiation

The ISLSCP "Initiative 1" data set (Sellers et al., 1995; Meeson et al., 1995) is arguably the most spatially and temporally comprehensive, uniform and contiguous data currently available for land-surface modeling studies. This data set includes many of the fields required to prescribe boundary conditions, initiate state variables, and force land system models. Data are available for 1987 and 1988, and most are spatially continuous over Earth's land surface on a $1^\circ \times 1^\circ$ equal-angle grid. The temporal frequency of the data set is monthly, but some important near-surface meteorological parameters are available as both 6-hourly and monthly means. Most of the near-surface meteorological variables were extracted from the operational forecast analysis archive of the European Center for Medium-Range Weather Forecasts.

In BATS, air temperature, dew point temperature, surface pressure, wind velocity, precipitation, and downward short-wave and long-wave radiation are used as forcing variables, and these variables are available as 6-hourly time series on the ISLSCP CDROM. They were interpolated

to provide hourly-mean values using a cubic spline interpolation method. The interpolated negative values of shortwave radiation were set to zero. The 6-hourly total precipitation was equally distributed as hourly values for that period. Saturated specific humidity was calculated from dew point temperature and surface pressure using Tetens formula (Riegel, 1974). Air temperature and dew point temperature were given at 2 m above the ground, while wind velocity was given at 10 m. For consistency, the values of wind velocity were extrapolated to 2 m assuming a logarithmic wind profile.

Although soil texture data are provided in the ISLSCP CDROM, soil color classes are not provided; hence (for consistency), both soil texture and the soil color used in the stand-alone BATS simulations were taken from the Wilson (1984) data set. Selection of a set of soil parameters corresponding to these soil classes was then made following Dickinson et al. (1986). In each grid square, surface soil temperature, deep soil temperature, canopy temperature and canopy air temperature were all initialized to the observed air temperature in the stand alone BATS model. Soil moisture in the upper, root, and total soil layer was initialized to a fixed fraction of available soil storage, which in most grid squares corresponds to 60-75% of saturation, depending upon soil textural class. The ground wetness factor was set to 1. Because BATS output can be sensitive to initial state variables, the model was "spun up" for 20 years (by cycling the observed forcing variables from first year of CDROM data set) before it was run forward to provide the results for 1987 and 1988 described below.

4.2 Model Results

Figure 10 illustrates the calculated latent heat flux (left) and sensible heat flux (right) given by the stand-alone BATS simulations, using parameter values defined for each grid square for the most common cover in Figures 10a, 10c, 10e, and 10g and using aggregation rules in Figures 10b, 10d, 10f, and 10h. The four panels at the top of the figure are average values calculated using forcing variables from the ISLSCP CDROM data set for June, July, and August 1988, while the four figures at the bottom of the page are the equivalent average values for December 1987 and January and February 1988. [Note: There was little obvious difference between fluxes calculated for equivalent seasons in 1987 and 1988 and, in comparison with the latent and sensible heat fluxes, the 3-month average soil heat flux calculated with both sets of parameters is everywhere very small in both winter and summer seasons.]

A distinctly regional behavior in terms of the calculated surface fluxes across the U.S. is illustrated in these figures. In summer, evaporation (Figure 10a-b) largely follows the distribution of precipitation across the country, and the eastern USA and gulf coast have significantly greater latent heat flux than the western USA. There is complementary behavior for sensible heat (Figures 10e-f) as might be expected from energy balance considerations. The east-west distinction persists in the winter season. Latent heat fluxes (Figure 10c-d) are again larger in the east than the west--although the distinction is less pronounced than in the summer, while sensible heat fluxes (Figures 10g-h) are calculated to be negative in the eastern USA but are positive and approximately equal to the latent heat flux in the western USA. The BATS simulated fluxes were also compared with estimated surface fluxes provided on the ISLSCP CDROM which were

extracted from the European Center for Medium-Range Weather Forecast (ECMWF) operational forecast analysis archive. The geographical patterns of those fluxes were broadly similar to those shown in Figure 10 in summer, although the ISLSCP data include some suspiciously high latent heat fluxes in portions of the eastern United States. The sensible heat flux reported on the ISLSCP CDROM also has zones of sensible heat which run east-west in the winter, rather than north-south.

The differences between the surface heat fluxes calculated with parameters for the most common cover minus those calculated with aggregate cover are shown in Figure 11. Those on the left of the page (Figures 11a-b) are for latent heat, and those on the right side of the page (Figures 11c-d) are for sensible heat. The difference in calculated fluxes is within $\pm 10\%$ in many locations and in both seasons. However there are significant differences in the sensible heat flux in the summer season (Figure 11c), which are particularly noticeable in parts of the western USA and are consistent with the large difference in surface albedo and leaf area index for the two parameterizations in this region. There is also an appreciable shift in the Bowen ratio in some parts of the eastern USA in winter months, with the aggregate parameters tending to give rather more latent heat but rather less sensible heat.

Figure 12 illustrates that the influence on calculated fluxes given by the two alternative parameterizations can differ greatly from one location to the next and that the different parameterizations can give persistent alteration in the calculated surface energy fluxes in specific locations. Figure 12a shows the monthly average latent heat and sensible heat fluxes for a grid cell in central Kansas, while Figure 12b gives a grid cell in southern Arizona. In Kansas, there is

little difference throughout the entire two-year period, but in Arizona including the influence of evergreen shrub land and semi-desert and, at higher elevations, evergreen trees, reduces the area-average albedo of aggregate cover and systematically increases the outgoing sensible heat flux because evaporation is water-limited in this region.

5. Summary and Discussion

Model experiments at the ABRACOS site (this study) and at the FIFE site (see Arain et al., 1996) using observations for validation and initiation, have given guidance on the aggregation rules appropriate to the BATS model. Fortunately, linear aggregation rules work well for most of the vegetation-related parameters, including (on the basis of results for the ABRACOS site) minimum stomatal resistance. Aerodynamic roughness length is the exception: in this case, a logarithmic aggregation rule is required (Table 1). Experiments conducted for the ABRACOS site were effective tests of aggregation rules for stomatal resistance, aerodynamic roughness length, and zero plane displacement because there were large contrasts in these variables between the forest and pasture patches. Mixtures of patches of tall and short crops are not uncommon in parts of the world where there has been human influence on land cover, and it is reassuring that aggregation rules for BATS' vegetation-related parameters can work within acceptable limits in this case.

However, both of the above-mentioned studies suggest that caution is necessary if aggregate representation is sought in a landscape where there are approximately equal areas with significant differences in plant-available soil moisture. Such differences may be man-made (by irrigation),

but the proportion of irrigated land within a GCM grid is usually small and its influence on the area-average fluxes likewise. Large spatial variability in plant-available moisture might have natural origin, perhaps resulting from intermittent, spatially heterogeneous rainfall. A parallel study (White et al., 1996) directly addressed this issue for the extreme case of convective rain falling in a semi-arid environment. White's study demonstrated that sufficiently large differences in soil moisture rarely, if ever, occur because the location of successive convective storms is random, and the size of individual storms is usually small compared to the amount of moisture that can be stored in the plants' rooting depth.

The current study reveals a natural situation where large spatial variability in plant-available soil moisture can compromise the use of an aggregate description of mixed vegetation cover. If patches of vegetation with markedly different rooting depths exist side by side, then prolonged rain-free periods can result in moisture shortage for the shorter-rooted vegetation, but not for the deeper-rooted vegetation. In the case of the Amazon river basin, such rain-free periods are not usually prolonged and, on the basis of the present study, their detrimental consequences when using aggregation equations is limited in any case. However, it is possible to envisage particular situations where this could be a more persistent problem. An example is the case of freely transpiring, riparian vegetation in a semi-arid environment growing alongside persistently flowing streams which are sustained by groundwater or distant mountain headwaters. In such extreme situations it is perhaps necessary to find a way to aggregate soil moisture and vegetation parameters in combination.

Notwithstanding the above comments, this study assumed that the BATS-specific aggregation rules given in Table 1 (with linear aggregation of stomatal resistance) are now sufficiently well-established to be applied in conjunction with remotely sensed land cover data. Accordingly, area-average values of BATS vegetation parameters were calculated for $2.8^\circ \times 2.8^\circ$ (CCM2) grids and $1^\circ \times 1^\circ$ (ISLSCP) grids across conterminous the United States using these aggregation rules with the USGS/EDC prototype land cover data set. The values of these aggregate-vegetation parameters were compared with those for the most common cover for each grid square. Significant differences were found in some key vegetation parameters such as aerodynamic roughness, minimum stomatal resistance, maximum and minimum leaf area indices, and albedo.

Stand-alone BATS model simulations were then performed using a 2-year-long time series of forcing variables from the ISLSCP CDROM data set. In general, the difference in fluxes calculated for aggregate cover and fluxes calculated for dominant cover was less than might have been anticipated on the grounds of the differences in individual parameters. In some cases, the fact that there are often co-located differences in two or more parameters helps to mitigate the net change in surface fluxes. Nonetheless, there were sizable regions of the USA where significant changes in calculated surface energy fluxes occurred in response to the different procedures for assigning vegetation parameters in the BATS model.

Further research is required to evaluate whether the difference in surface fluxes given by using aggregate (as opposed to most-common cover) parameters has a significant influence on modeled climates given by GCMs. However, global application of aggregation rules to improve GCM performance (in conjunction with upcoming remotely sensed AVHRR Path-Finder and EOS

MODIS data sets) is at this stage arguably more profoundly limited by shortcomings in the validation of the cover class-specific values of vegetation parameters used, for example, in BATS. There is an urgent need to make effective use of the data now available from field sites around the world (for evergreen needleleaf forest, savannah grasslands, semi-desert, marsh land, short grass, and crops) to define the value of parameters for individual cover classes.

Acknowledgments

We are grateful to Robert Dickinson and Eleanor Blyth, who respectively provided us with the BATS and the ABL model, and to John Gash and Tom Loveland for respectively providing the ABRACOS and the USGS/EDC Land cover data sets. We also acknowledge many scientists who acquired and distributed the USGS-EROS land cover and the ABRACOS data sets and, in particular, Ivan Wright, for his help and advice during this study. We would also like to thank Jean Morrill, Paul Houser, Cary White, Dan Braithwaite, Jaime Garatuza, James Broermann, Jim Washburne, Corrie Thies, and Soroosh Sorooshian for their help and encouragement throughout. The research described in this paper was carried out under NASA grant number NAGW-3368#1. Altaf M. Arain was partially supported by the Ministry of Science and Technology, Pakistan and NASA Global Change Research Fellowship.

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Figure Captions

Figure 1.

(a) and (b) show a comparison between sensible heat and latent heat fluxes (shown as dotted lines and full lines respectively) at the ABRACOS pasture site for July 21, 1991. The lines with symbols are observations, while lines without symbols are model-simulated fluxes with minimum stomatal resistance of (a) 200 m^{-1} and (b) 40 m^{-1} , respectively. (c) and (d) show net radiation and modeled latent heat fluxes (shown as broken lines and full lines, respectively) at the ABRACOS forest site (c) for July 21, 1991, which was a day without moisture shortage, and (d) for October 30, 1990, which was a day with significant moisture shortage. The lines without symbols are modeled fluxes with 1.5m deep roots mainly concentrated near the surface, while lines with symbols are with 4m deep roots more equitably distributed.

Figure 2.

Comparison of surface fluxes of (a) sensible heat, (b) latent heat, (c) soil heat, and (d) momentum at 10:00 a.m. on October 30, 1990 displayed as a function of position across the 2000-m wide modeled domain used in the BATS-ABL model. In each case, the full line is the flux calculated with representation of distinct patches of pasture and forest, with the forest patch lying between 500 and 1500 m, and the broken line is the average value of these same fluxes across the domain. The dotted line is the flux calculated with aggregate vegetation covering the entire domain.

Figure 3.

Comparison of area-average fluxes of (a) sensible heat, (b) latent heat, (c) soil heat, and (d) momentum calculated by the BATS-ABL model for the ABRACOS site with aggregate representation and explicit representation for patches of forest and pasture. These calculations were made at three times of day on five days using BATS vegetation parameters which had been slightly revised to enhance the agreement between observed and model-calculated fluxes.

Figure 4.

Same as Figure 4 but using the original, unmodified BATS vegetation parameters.

Figure 5.

(a) Calculated latent heat, sensible heat given by stand-alone BATS models for the period, October 14, 1990 to November 1, 1990 when there were dry soils. In each case, the full line is the area-average flux with separate modeling of equal areas of forest and pasture, while the broken line is the aggregate representation of these two covers. Lines without symbols are for latent heat, while those with symbols are sensible heat. (b) is similar to (a) except calculations are made for the period 12 January, 1991 to 31 January 1991 when there were moist soils. (c) Soil moisture wilting factors calculated by the stand-alone BATS model for the complete study period from October 4, 1990 to September 20, 1991 for upper and root zone soil layers under both pasture and forest.

Figure 6.

Comparison of the calculated fluxes given with three alternative aggregation rules for minimum stomatal resistance on (a) October 30, 1990 (a day with very dry soils), and (b) August 28, 1991 (a day with wet soils). Calculated latent heats are shown with the lines without symbols, and sensible heat fluxes are shown with lines with crosses.

Figure 7.

(a) The number of surface covers which cover more than 5% in each $2.8^\circ \times 2.8^\circ$ grid square. (b) The fractional area covered by the most common remotely sensed cover dominant class in each $2.8^\circ \times 2.8^\circ$ (CCM2) grid square.

Figure 8.

Percentage difference in BATS-relevant vegetation parameters derived from the USGS/EDC data set using aggregation rules applied to $2.8^\circ \times 2.8^\circ$ (CCM2) grid squares over the conterminous United States relative to the equivalent value for the single most common BATS cover class in each grid square. The separate figures respectively correspond to the following parameters: (a) vegetation cover fraction; (b) minimum stomatal resistance; (c) albedo (average of visible and near infrared range); (d) warm season (maximum) leaf area index; (e) cool season (minimum) leaf area index; and (f) aerodynamic roughness length.

Figure 9.

Same as Figure 8 but for $1^\circ \times 1^\circ$ (ISLSCP) grid squares.

Figure 10.

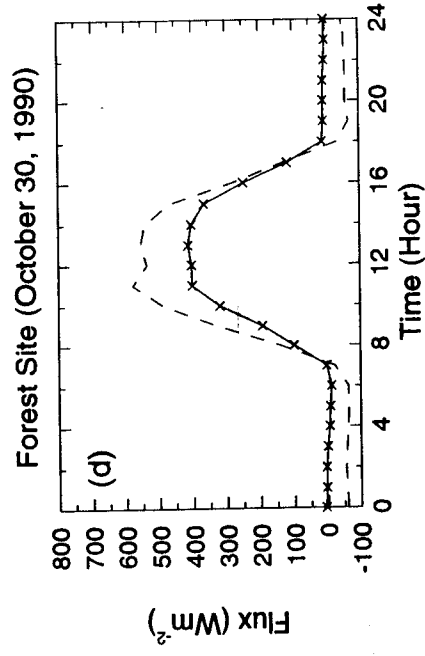
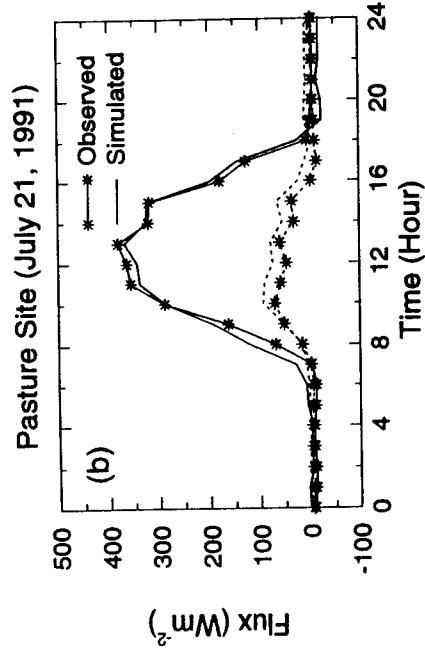
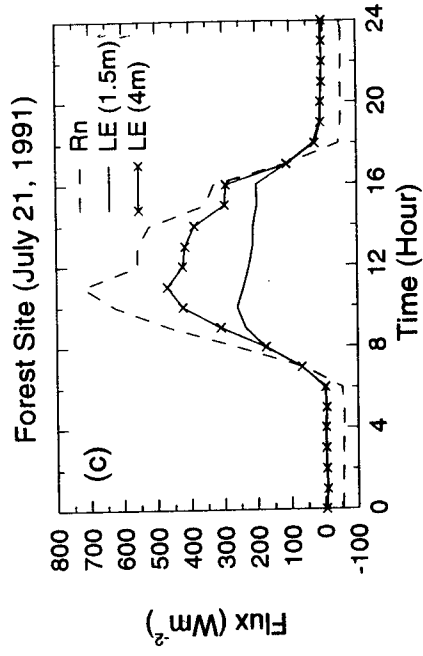
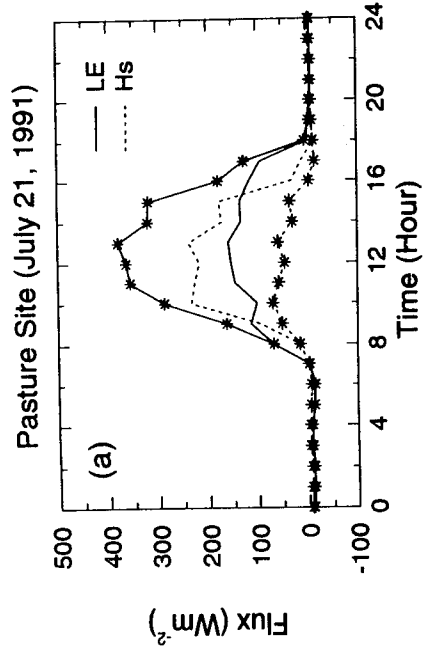
Calculated values of surface heat fluxes given by applying the ISLSCP forcing data to a two-dimensional array of stand-alone BATS models spanning the conterminous United States with parameters for the single most common cover class [(a), (c), (e), and (g)] and for aggregate-cover [(b), (d), (f), and (h)]. The figures on the left-hand-side [(a), (b), (c), and (d)] are for latent heat, while the figures on the right-hand-side [(e), (f), (g) and (h)] are for sensible heat. The top four figures [(a), (b), (e), and (f)] are average values for the months of June, July and August 1988, while the bottom four figures [(c), (d), (g), and (h)] are average values for December 1987 and January and February 1988.

Figure 11.

Difference between the calculated values of surface heat fluxes shown in Figure 10 using parameters for most common cover minus those parameters for aggregate cover. The figures on the left-hand-side [(a) and (b)] are for latent heat, while the figures on the right-hand-side [(c) and (d)] are for sensible heat. The top two figures [(a) and (c)] are average values for the months of June, July, and August 1988, while the bottom two figures [(b) and (d)] are average values for December 1987 and January and February 1988.

Figure 12

Monthly average latent heat and sensible heat fluxes for 1987 and 1988 calculated with a stand-alone BATS model forced with ISLSCP CDROM data using aggregate-cover parameters (shown with a symbol) and most common cover parameters (shown without symbol) for a $1^\circ \times 1^\circ$ ISLSCP grid square (a) in Kansas and (b) in southern Arizona.



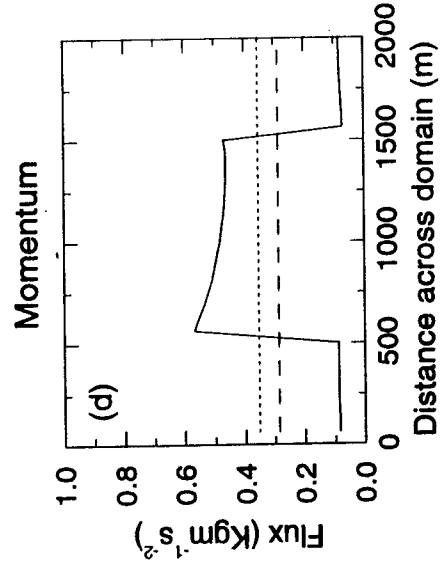
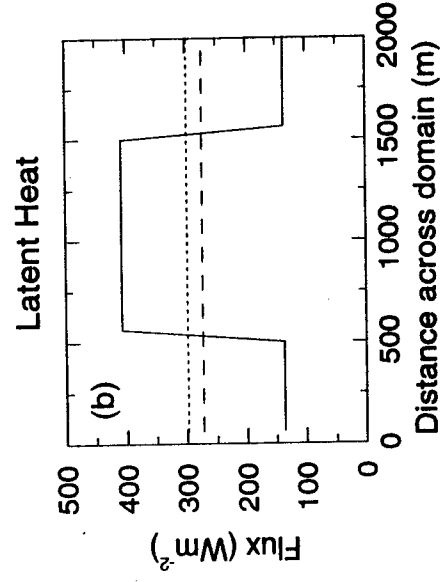
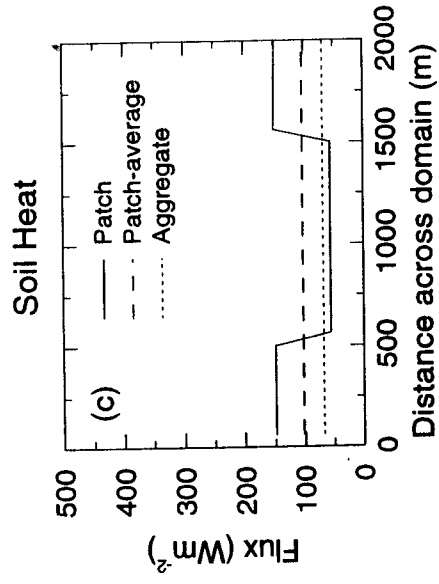
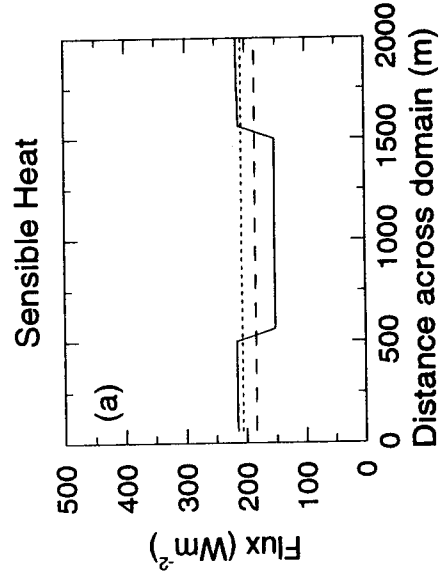
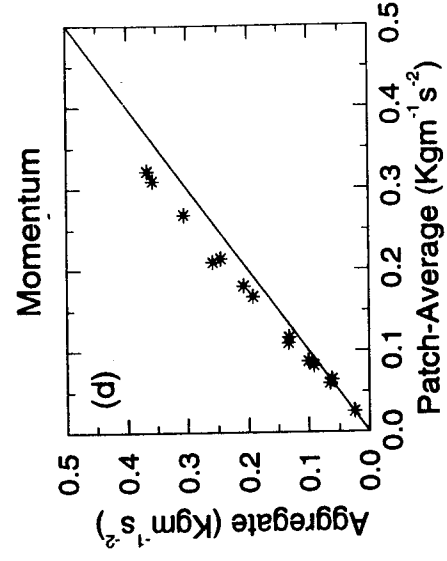
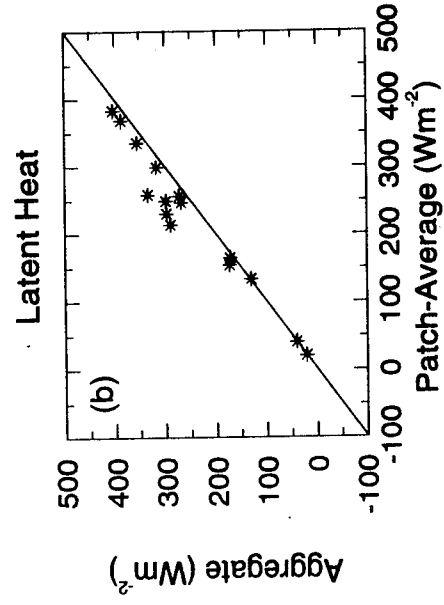
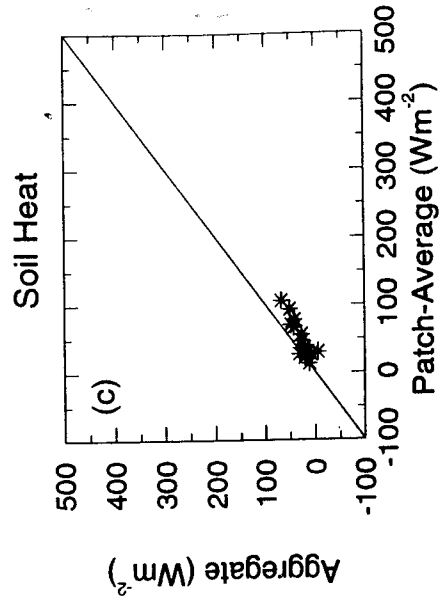
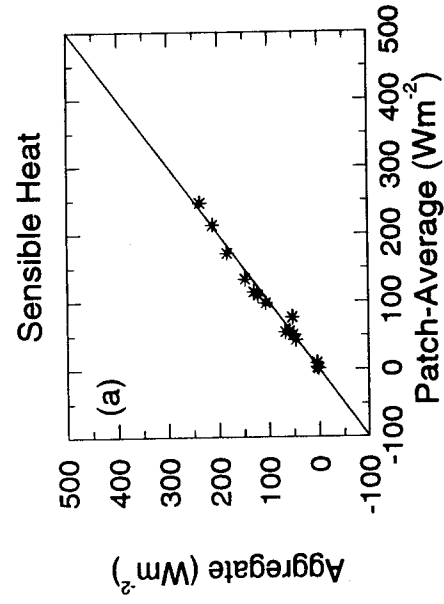
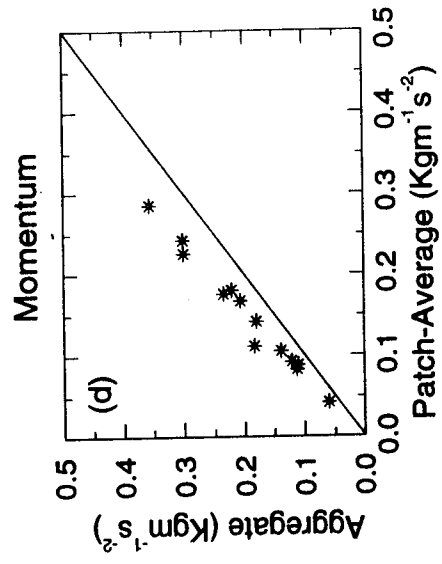
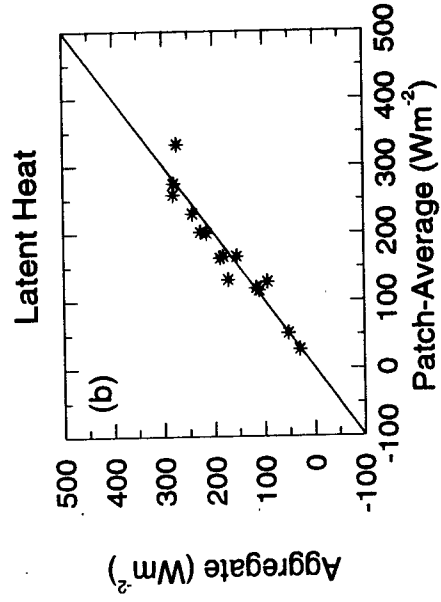
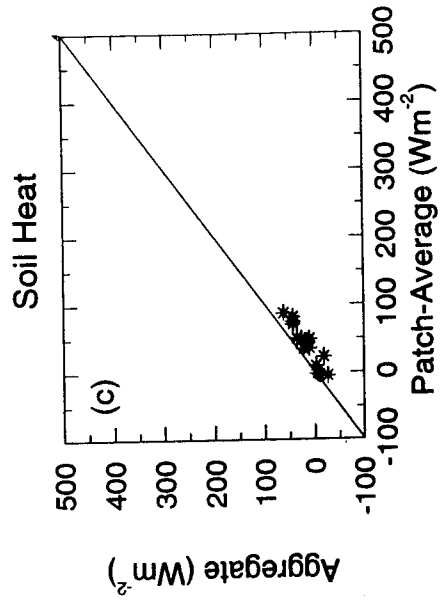
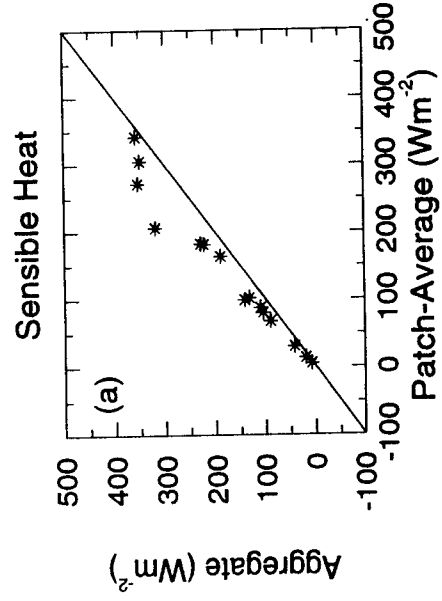
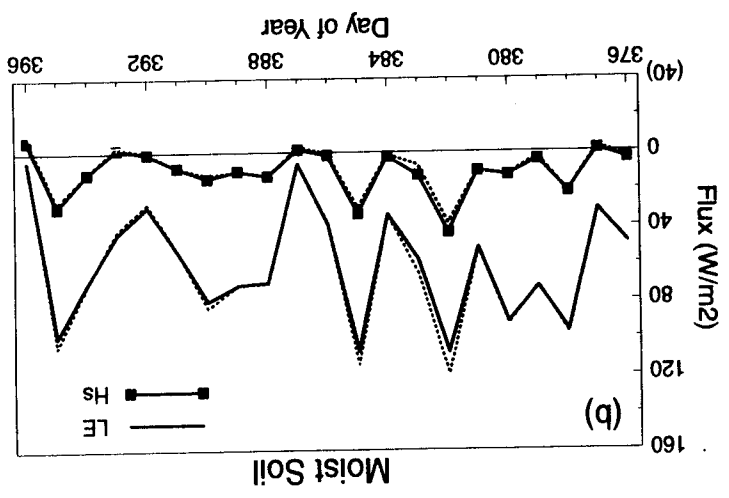
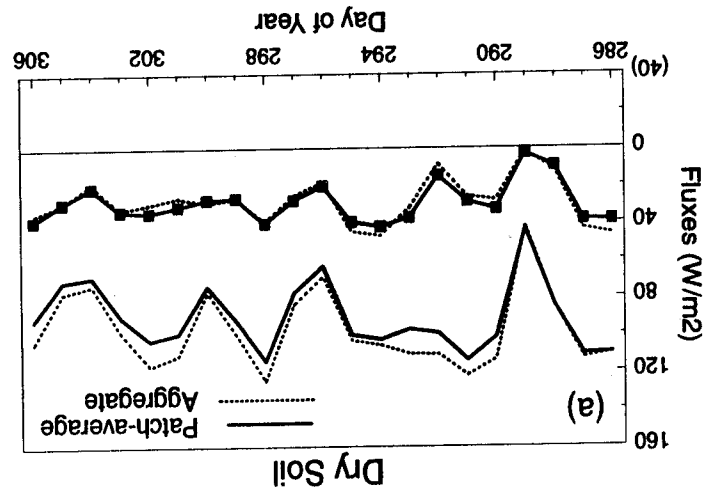
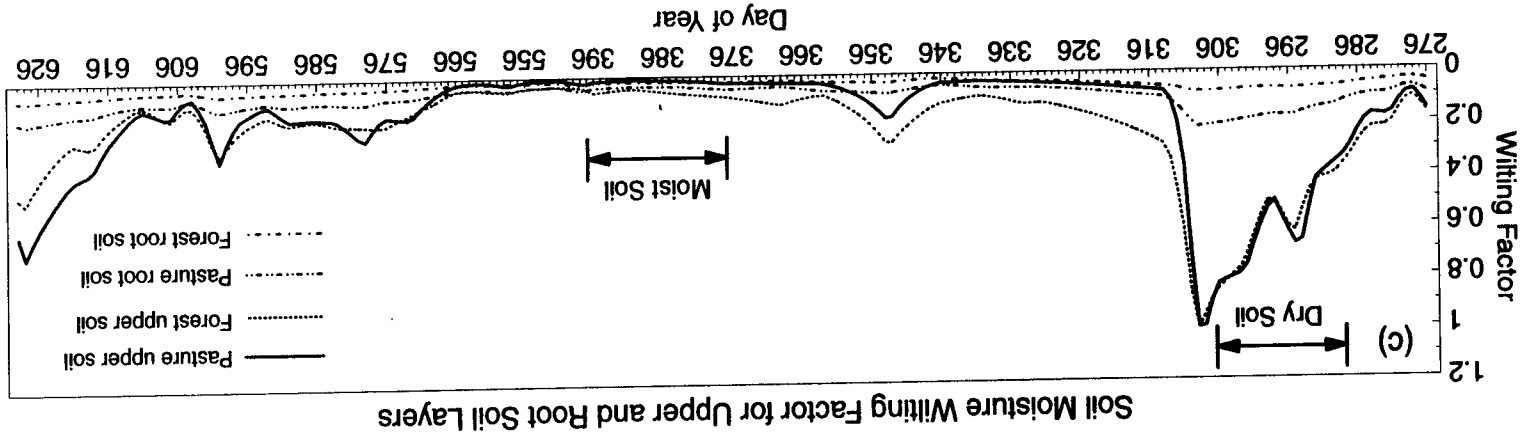
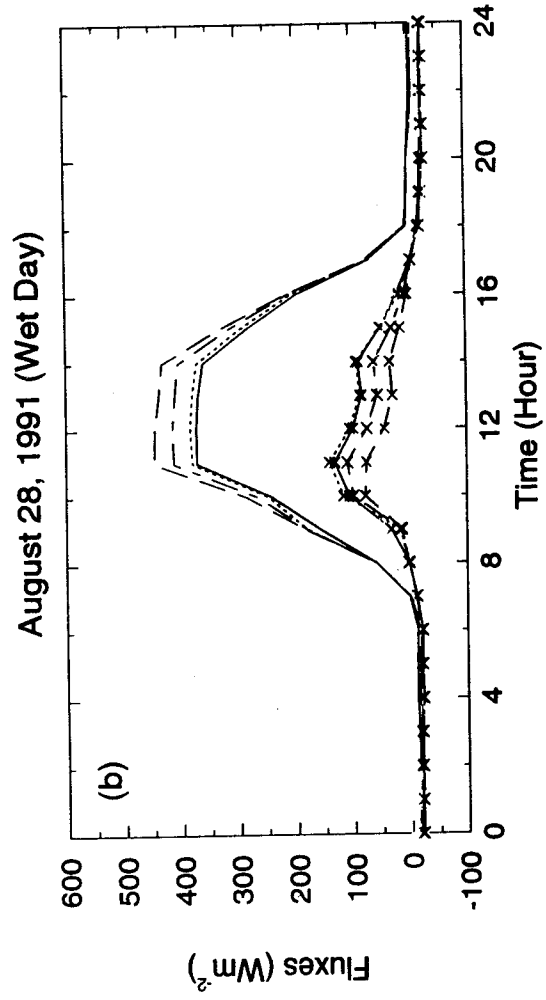
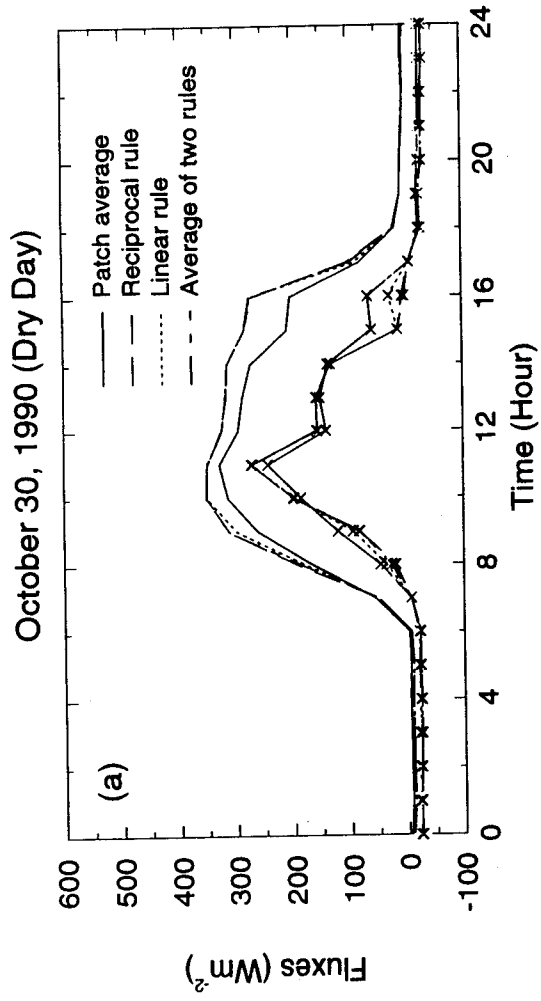


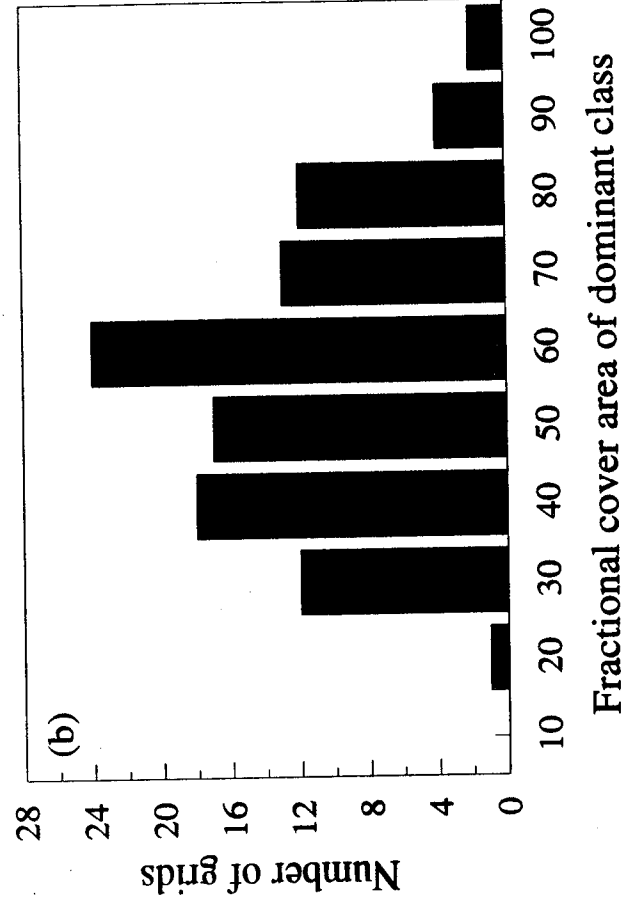
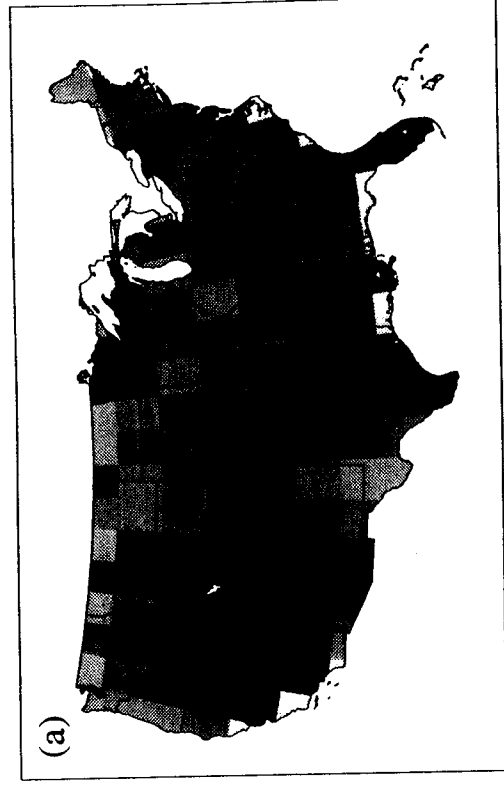
Figure 2

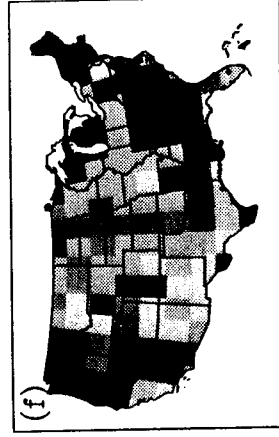
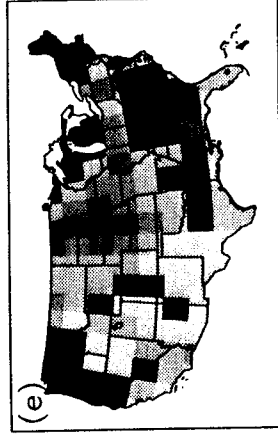
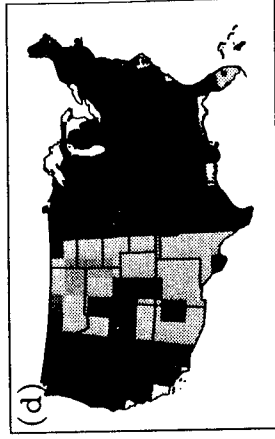
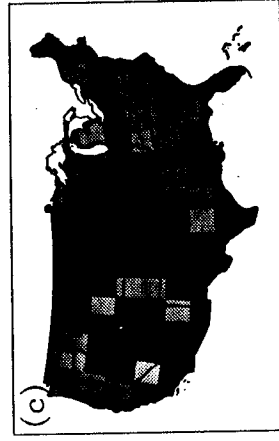
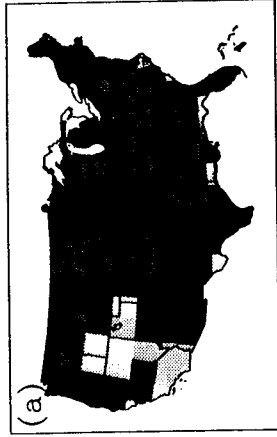


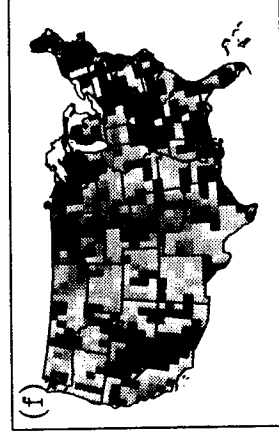
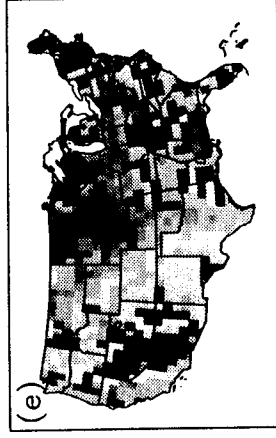
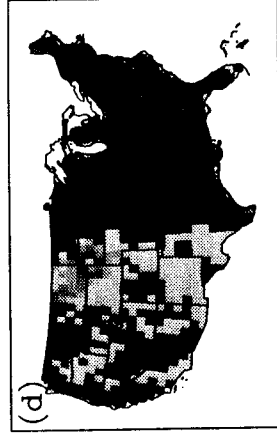
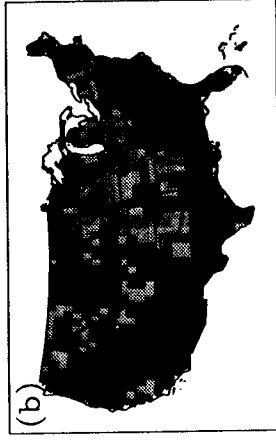












Latent Heat

Summer Dominant



Summer Aggregate



Winter Dominant



Winter Aggregate



Sensible Heat

Summer Dominant



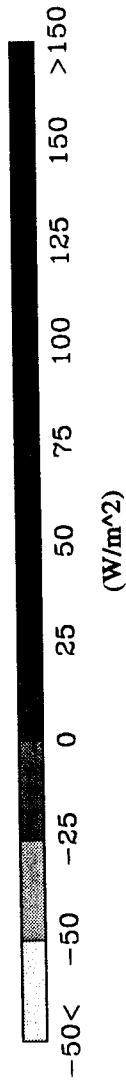
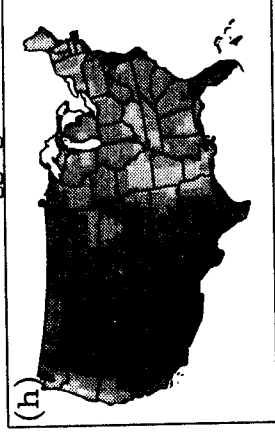
Summer Aggregate



Winter Dominant



Winter Aggregate



Latent Heat
Summer



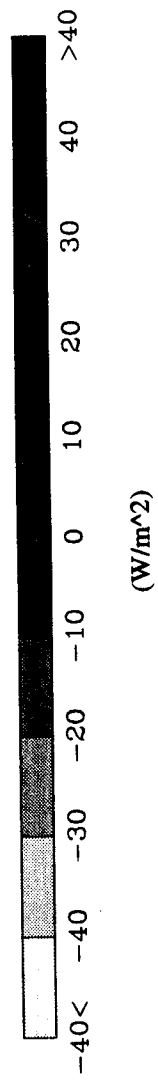
Sensible Heat
Summer



Winter



Winter



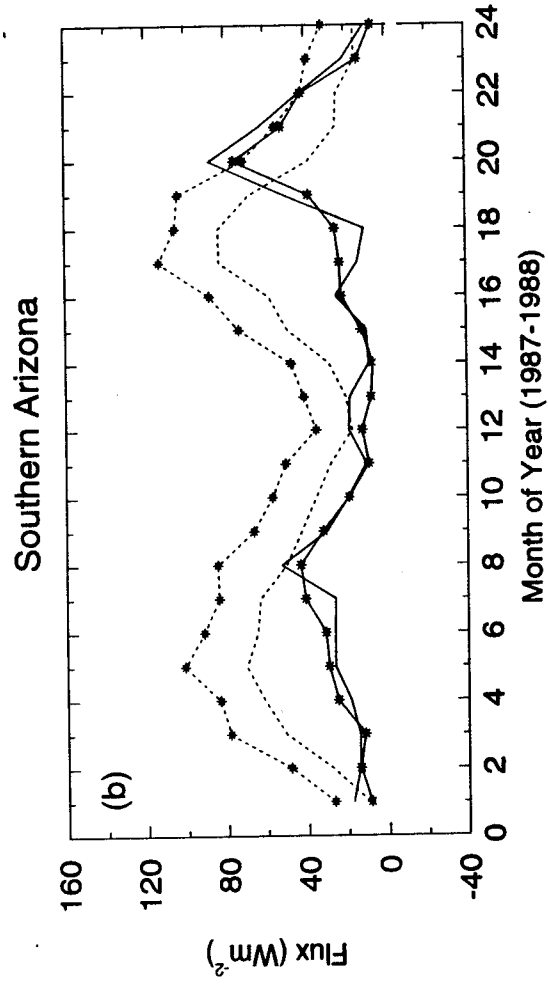
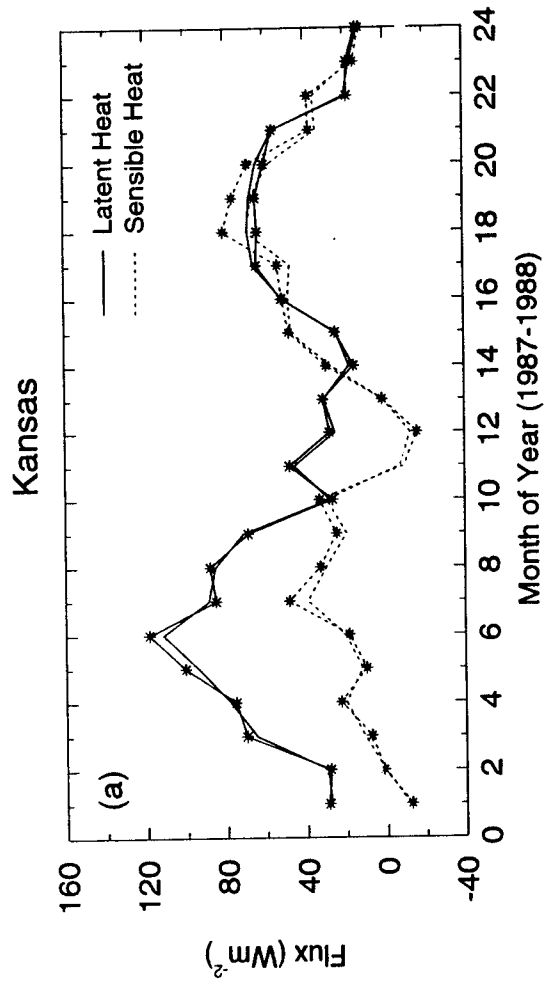


Figure 12

Table 1. Aggregation rules for BATs vegetation parameters

Parameter	Aggregation Rule	Comment
Zero plane displacement height (d)	$\bar{d} = \sum w_i d_i$	Linear effect on surface processes
Minimum stomatal resistance (r_{smin})	$\overline{r_{smin}} = \sum w_i r_{smin_i}$ $\frac{1}{\overline{r_{smin}}} = \sum \frac{r_{smin_i}}{w_i}$ $\overline{r_{smin}} = \overline{r_{smin}} + \frac{1}{\overline{r_{smin}}}$	Linear rule Reciprocal rule Average of linear and reciprocal rules
Roughness length for momentum (z_0)	$\overline{\ln \left(\frac{l_p}{b} \right) \frac{z_0}{2}} = \sum w_i \ln \left(\frac{l_p}{b} \right) \frac{z_{0i}}{2}$	Blending height given as $l_p = 2(u_* / V_a)^2 l_c$ where l_c , u_* , and V_a are the horizontal length scale, friction velocity, and wind flow at blending height, respectively

The subscript “i” represents cover (vegetation) type, and “w_i” represents the fractional area of that cover type. Parameters (see Table 2 for a list) that use a linear area-weighted average have been omitted for brevity, with the exception of displacement height *d*, which is included as an example. Three alternate rules are given for combining minimum stomatal resistance. On the basis of tests made at the ABRACOS site (see Section 2), the linear averaging rule is recommended and was adopted in this study.

Table 2. Values of the BATs vegetation parameters for the ABRACOS-relevant cover types

Parameter			
Short Grass	Maximum fractional vegetation cover	0.8	0.9
	Factor for seasonal change in vegetation cover	0.1	0.5
Short Grass	Roughness length (m)	0.02	2.0
	Depth of the rooting zone (m)	1	1.5 (4.0)
Short Grass	Depth of the upper soil layer (m)	0.1	0.1
	Fraction of water extracted by upper layer roots	0.8	0.8 (0.2)
Short Grass	Albedo for wavelength < 0.7 μm	0.1	0.04
	Albedo for wavelength > 0.7 μm	0.3	0.2
Short Grass	Minimum stomatal resistance (sm^{-1})	200 (40)	150
	Maximum leaf area index	2	6
Short Grass	Minimum leaf area index	0.5	5
	Stem and dead matter area index	4	2
Short Grass	Inverse square root of leaf dimension ($\text{m}^{-1/2}$)	5	5
	Light sensitivity factor (m^2W^{-1})	0.02	0.06
Short Grass	Zero plane displacement (m)	0	18

* Modified parameter values are in parentheses

Evergreen Broadleaf Tree

Table 3. BATS soil parameters used in ABRACOS study

(a) Soil texture parameters for class 10

Parameter	Value
Porosity	0.60
Minimum soil suction (mm)	200
Saturated hydraulic conductivity (mm s^{-1})	0.0016
Ratio of saturated thermal conductivity to that of loam	0.8
Exponent B	9.2
Moisture content relative to saturation at which transpiration ceases	0.487

(b) Albedo for soil color class 7

Condition	Albedo
Dry soil, wavelength $< 0.7 \mu\text{m}$	0.12
Dry soil, wavelength $> 0.7 \mu\text{m}$	0.24
Wet soil, wavelength $< 0.7 \mu\text{m}$	0.06
Wet soil, wavelength $> 0.7 \mu\text{m}$	0.12

Table 4. BATS state variables, driving variables, and parameters

States	
Surface soil temperature (°K)	
Subsurface temperature (°K)	
Temperature of foliage (°K)	
Temperature of air within foliage (°K)	
Water in upper soil layer (mm)	
Water in root zone soil (mm)	
Total water in soil (mm)	
Ground wetness factor (fraction)	
Dew on leaves (mm)	
Driving Variables (Boundary Inputs)	
Incident solar radiation (Wm^{-2})	
Wind velocity (ms^{-1})	
Air temperature at interface level (°K)	
Specific humidity at interface level (gKg^{-1})	
Precipitation (mm)	
Surface pressure (mbar)	
Downward longwave radiation (Wm^{-2})	
Parameters	
Interface height (m)	
Vegetation parameters (see Table 2)	
Land or sea flag	
Soil texture parameters (see Table 3)	
Soil color parameters (see Table 3)	

Table 5. Statistics for BATS validation at pasture site

(a) $rs_{min} = 200 \text{ sm}^{-1}$

Flux	Mean Bias (Wm^{-2})	RMSE (Wm^{-2})	Correlation Coefficient
Latent heat	-38	82	0.90
Sensible heat	+18	57	0.85
Soil heat	-2	35	0.88
Net radiation	-31	42	0.99

(b) $rs_{min} = 40 \text{ sm}^{-1}$

Flux	Mean Bias (Wm^{-2})	RMSE (Wm^{-2})	Correlation Coefficient
Latent heat	-11	54	0.91
Sensible heat	-4	40	0.80
Soil heat	-1	33	0.89
Net radiation	-24	36	0.99

Table 6. Days and times for which aggregate and patch simulations were performed and observed soil moisture at corresponding mid-day time.

Day of Year	Time of day (hr)			Moisture in 0.1m Soil Layer (mm)	Moisture in 1m Soil Layer (mm)
October 26, 1990	09:00	11:00	16:00	32	352
October 30, 1990	08:00	10:00	13:00	30	345
December 10, 1990	08:00	11:00	16:00	41	402
July 3, 1991	08:00	12:00	15:00	44	408
August 28, 1991	09:00	12:00	15:00	40	391

Executive Summary of the *Tucson Aggregation Workshop*

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Abstract

"*Aggregation*" refers to spatial averaging of some heterogeneous surface variable to obtain an effective value representative of an area. The effect of surface heterogeneity on interactions between land and atmosphere is relevant to near-surface hydrology, ecology, and climate, and is the common theme of the papers in this issue. Even though the full effect of heterogeneity must be neglected due to limited spatial resolution of large-scale models, it is important to understand when and how the presence of heterogeneity requires recognition in any aggregate representation. In March of 1994, a workshop, which has come to be known as the "*Tucson Aggregation Workshop*", was convened to assess the state-of-the-art in aggregation research, and the papers in this issue are the product of that workshop. The principle findings of the workshop can be summarized as follows: (a) Substantial progress has been made in producing aggregated representations of flat terrain. Simple *aggregation rules* applied to surface properties have given rise--in some studies--to simulated surface energy fluxes that are within 10% of fluxes produced from models with full representation of heterogeneity. (b) Aggregation

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rules are relatively straightforward in the case of patch-scale heterogeneity (variability on the order of 100s-1000s of meters) of vegetation characteristics which control surface exchanges, although aggregation of soil hydraulic properties and possibly soil moisture remains problematic. In addition, some of the effects of meso-scale heterogeneity (variability on the order of 10-100 km) in surface cover will need to be addressed through more complicated types of parameterization. (c) There is convincing evidence that the regional energy balance (over, say, 10^5 km^2) is insensitive to gentle topography, providing that surface vegetation and water availability are uniform, but in mountainous terrain the influence of topography on near-surface meteorology most likely needs to be considered. (d) It appears that the value of simple combinations of remotely sensed radiances representing areal-average measurements are only slightly influenced by unresolved variability, although the averaging of some derived variables based on these radiances offers a greater challenge, especially with sparse canopies.

1. INTRODUCTION

The effect of land surface heterogeneity on the atmosphere and on the surface energy balance is a topic that has attracted widespread interest because understanding this effect is fundamental to a comprehensive knowledge of regional and global hydrometeorological processes. Moreover, many investigators are concerned that inadequate treatment of heterogeneity may weaken our confidence in large-scale models which do not resolve heterogeneity at scales smaller than the model grid. Interest in heterogeneity has been further spurred by several technical advances, not the least of which is the availability of satellite data. Remote sensing technology offers high-resolution data which can be used to quantify regional

and global heterogeneity and make area-average measurements representing the effective area-average value of surface parameters. Computational advances and increased interest in climate, and therefore in the modeling of land-atmosphere interactions, have also promoted interest in surface heterogeneity.

"*Aggregation*" generally refers to spatial averaging of some heterogeneous surface variable such as albedo, soil hydraulic properties, soil moisture, fraction of vegetation cover, surface temperature, surface reflectance, sensible heat flux, latent heat flux, surface resistance, aerodynamic resistance, or aspects of topography; or it refers to spatial averaging of some near-surface meteorological field such as temperature, humidity or precipitation. There is the question of how to "average" (arithmetic or logarithmic being a few of the possibilities), and the size of the region over which averaging should be performed. This size depends on the degree of heterogeneity and whether there is a causal relationship between the variable being averaged and the quantity to be calculated in the model. There is a possibility that aggregation will fail when a heterogeneous variable has a non-linear relationship with some other variable of interest. Moreover, aggregation strategies may be dependent on model formulation. Aggregation is a more limited enterprise than "*scaling*", because scaling seeks to find a basis for relating a phenomenon at one scale to an analogous phenomenon at other scales. However, several papers in this issue do address scaling.

Interest in the "aggregation problem" is largely motivated either by the desire to make efficient use of highly resolved spatial data, or by the desire to proceed confidently without utilization of detailed data. In other words, it seeks to address the question, "How can we model spatially variable processes using a grid size which is coarse enough to be economical, yet fine

enough that results are not affected by sub-grid scale variability?" However, the topic of aggregation is equally pertinent to the question of adequate spatial resolution of measurements; hence there is a need to investigate the effect of spatial resolution on the accuracy of remotely sensed measurements.

In March of 1994, a workshop was convened in Tucson, Arizona to assess the state-of-the-art in aggregation research. The workshop was jointly sponsored by the *Biological Aspects of Hydrologic Cycle* (BAHC) Core Project of the International Geosphere-Biosphere Program and the *International Satellite Land Surface Climatology Project* (ISLSCP) of the World Climate Research Program. In some part, the organization of the workshop, and in major part the subsequent publication of this Special Issue was also supported by the U.S. National Aeronautic and Space Administration (NASA), under its *Mission to Planet Earth's* Terrestrial Ecology and Modeling Program.

Sessions were organized around the topics of patch-scale land-atmosphere interactions (variability on the order of 10s-100s of meters), meso-scale land-atmosphere interactions (variability on the order of 10-100 km), land-atmosphere interactions in rugged terrain, the aggregation of remotely sensed variables, and soil hydrology. Each session consisted of invited presentations by leading researchers, followed by a guided discussion. As organizers and rapporteurs of the workshop, we are grateful to the speakers and their colleagues who have prepared the written papers which resulted from those presentations. The discussions which followed each presentation and each session have in different ways been incorporated into the written papers. Some of the printed papers conclude with a separate, edited section containing the recorded comments or questions posed by workshop participants, and the speaker's response

to these. In other cases, the authors chose to respond to the audience's questions by addressing these in the main body of the paper they subsequently prepared. It is our opinion that the ensuing suite of papers comprise a focused but broad overview of our current understanding of issues related to the aggregate representation of land-atmosphere interactions, and it is with satisfaction and some pride that we propagate them in this Special Issue of the *Journal of Hydrology*.

2. SUMMARY OF WORKSHOP FINDINGS

The principal findings of the workshop can be summarized as follows.

- Aggregation of land surface properties appears to be successful to within an accuracy of about 10% in many--but not all--circumstances. Stated more precisely, effective parameter values representing the areal averages of land surface properties in models of surface-atmosphere interactions have been successfully calculated from simple averaging rules, with the form of the latter being related to the nature of the variable being averaged (e.g., Shuttleworth, 1991). Patch-scale and meso-scale simulations show that energy fluxes calculated from these effective (aggregated) parameters can be within 10% of energy fluxes obtained from higher resolution simulations (Dolman and Blyth, this issue; Sellers *et al.*, this issue; Noilhan *et al.*, this issue).

Using a combination of wind tunnel experiments, theoretical analysis, and simulation, Raupach and Finnigan (this issue) showed that the regional energy balance is insensitive to the presence of hills of moderate size, providing that the nature of the vegetation and soil at the surface and the soil water available to the vegetation are uniform. Aggregation of near-surface meteorology considered in isolation is likely to be successful for slopes up to 20%.

The above successes are encouraging, but additional work is needed in (a) the aggregation of soil hydraulic properties; (b) lateral near-surface water and groundwater flow, and (c) examination of the effect of distinct lateral changes in vegetation height. In addition, some--but not all--researchers point to the need for additional work in the aggregation of soil moisture. While there has been substantial progress in understanding scaling of ecohydrologically relevant soil parameters at plot and field scales (1-10000 m²) (Kabat *et al.*, *this issue*), this progress has been little recognized by the large-scale meteorological modeling community; the applicability of these scaling procedures at large scales remains underexplored. In terms of soil moisture, several researchers (Sellers *et al.*, *this issue*; Wood, *this issue*) have shown that neglecting small-scale moisture variability may compromise coarse-grid simulations of areal-average evaporation, though Sellers *et al.* (*this issue*) view this as of secondary significance.

- Meso-scale heterogeneity in land surface properties is now known to be capable of generating meso-scale circulations which can have a significant effect on vertical energy transfers within the atmosphere. Parameterization of this phenomenon, which would allow General Circulation Models to accommodate these additional sub-grid scale atmospheric transport processes, is a topic of active research (Pielke *et al.*, *this issue*). However, some researchers (Noilhan *et al.*, *this issue*) view the need to provide such parameterization with less urgency, drawing attention to the moderating effect of winds.

- The purpose of many aggregation studies is to provide information which is intended to refine or stimulate regional and global models of the interactions between soil, vegetation, energy, and water. The basic tools for regional ecohydrological modeling have already been developed and applied in mountainous terrain (Thornton *et al.*, this issue). Adequate specification of finely resolved near-surface meteorology, particularly precipitation, is one of the difficulties that needs to be addressed, but there is currently no universally accepted procedure for doing this.
- Remotely sensed vegetation indices contain useful information on the bulk stomatal resistance and photosynthetic uptake of vegetation (Sellers *et al.*, 1992), but the roles of vegetation type and nutrition on the interpretation of these indices requires further investigation.
- Aggregation of remotely sensed measurements in sparse canopies can be accomplished with little error in some circumstances (such as aggregation of surface temperature from 1 m² to 1 km²) but not others (such as aggregation of sensible heat to 1 km²) (Moran *et al.*, this issue).

3. Summary of Progress in Aggregation Research and Recommendations for Future

Research

During the final discussion period, workshop speakers and participants evaluated the significance of progress in aggregation research made during the last eight years. A consensus emerged as follows.

- Substantial progress has been made in modeling the interaction between water, energy, vegetation, and soils on scales ranging from patch to global. In many ways, the purpose of future aggregation studies is to add refinements to this past comprehensive endeavor.
- There has been significant and satisfactory progress in developing aggregation techniques to deal with vegetation-dependent surface properties such as albedo, aerodynamic resistance, surface resistance, and roughness length.
- There has been only limited progress in defining aggregation rules for subsurface variables which are practical at the regional scale, though there has been some progress in coupling subsurface flow models to models of surface energy and water balance.
- Substantial progress has been made in the area of aggregation of remotely sensed variables, and it is now recognized that surface temperature and spectral vegetation indices can be easily aggregated at scales ranging from meters to kilometers. However, it is also recognized that there can be significant differences with respect to aggregation strategies for

sparse and dense canopies for variables derived from remotely sensed data. We do, however, have some knowledge of the factors that may introduce troublesome non-linearities into the remotely sensed measurements. In order of decreasing importance, these are soil and soil moisture, vegetation, and topography.

Workshop participants also discussed the most promising avenues of future research. This discussion can be summarized as follows.

- The quantification of fine-scale heterogeneity (of soils, fluxes, and remotely sensed variables) remains a clear and high priority;
- Most participants favored a "broad-front" approach combining bold application of current aggregation strategies at global and regional scales and continuing mechanistic research into specific aggregation issues;
- The topic of soil moisture and heterogeneity of soil properties clearly requires additional research, and the role of nutrition in controlling soil-vegetation-atmosphere interactions should not be overlooked; and
- Finally, there is little question that there is a continuing need for field observations over a range of scales, for validation of models, aggregation schemes, and scaling relationships. These observations should preferably encompass seasonal and intra-annual time scales.

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Acknowledgments. Our assessment of workshop findings and recommendations is based on the papers and discussions which took place at the Tucson Aggregation Workshop, held in Tucson, Arizona, March 23-24, 1994. The contribution of workshop participants (approaching 100, far to numerous too mention by name) and of workshop speakers (Han Dolman, Karen Humes, Pavel Kabat, Susan Moran, Joel Noilhan, Roger Pielke, Mike Raupach, Steve Running, Piers Sellers, and Eric Wood) to the content of this summary was therefore fundamental to its writing. Financial and administrative support for the Tucson Aggregation Workshop was provided by the *Biological Aspects of Hydrologic Cycle* (BAHC) Core Project of the International Geosphere-Biosphere Program (IGBP) and the *International Satellite Land Surface Climatology Program* (ISLSCP) of the International World Climate Program. The workshop and, in major part, the subsequent publication of this Special Issue was further substantially supported by the U.S. National Aeronautics and Space Administration (2485-MD/BGE-0011), under its Terrestrial Ecology and Modeling Program, which is under the direction of Diane Wickland.

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SOIL - VEGETATION - ATMOSPHERE RELATIONS: PROCESS AND PROSPECT

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1. Introduction

The driving mechanism of the climate system and the primary instrument of climate change is the capture of solar energy, and the way this energy is shared within the atmosphere. On average about half of the sun's energy reaches the earth's surface. From there it returns to the atmosphere in different forms - as reflected or emitted radiation, or as latent or sensible heat, depending on the nature of the underlying surface (Verstraete and Dickinson, 1986). The way in which such solar energy is shared is determined in substantial measure by the extent and availability of water at or near the surface.

Taking the global average, only about 40% of the precipitation measured as falling on continental surfaces is then measured to run off to the oceans (UNESCO, 1978) - which implies that there is "double counting" of precipitation, *i.e.* liquid or solid water falls, is re-evaporated, and falls again. The extent to which such recycling occurs is influenced by soil-vegetation-atmosphere processes, and by the structure and the level of activity of the biosphere. The size and shape of vegetation influences the amount of solar radiation, momentum and rain water captured at the ground, and the depth to which soil water is accessible to the atmosphere via roots. The level of activity of vegetation controls the rate of water loss directly via stomatal control on transpiration rate; it also has more subtle influence, by changing the hydrological properties of the soil to alter infiltration and bulk soil porosity.

The nature of the earth's surface - whether it is water or land, and what vegetation covers

Layer" up to a few meters above the ground. Within this layer the turbulent mechanism is usually a mixture of forced convection, which is determined by wind and by the aerodynamic roughness of the underlying vegetation and soil; and convection, which is influenced by the amount of energy leaving the ground as heat. Models of surface energy exchange must necessarily include simultaneous description of momentum exchange, since this determines forced convection; models often include some iteration, i.e. they first-compute energy fluxes with convection, then include this process as a correction using the "first guess" estimate of sensible heat.

Very commonly the movement of energy, momentum and other atmospheric constituents (carbon dioxide) by diffusion processes is represented in soil-vegetation-atmosphere models by use of resistance analogies. In general, the diffusion equations are in the form so that fluxes of energy, for example, are proportional to differences in energy potentials (vapor pressure for latent heat; temperature for sensible heat), divided by a resistance (see Shuttleworth, 1991 for details). Some more complex models will include a resistance to represent exchange by molecular diffusion through the boundary layer adjacent to transpiring leaves and soil, but more commonly they will include a resistance, which describes diffusion through the stomata from inside to outside and an "aerodynamic resistance", which describes diffusion from within the vegetation canopy to a prescribed height in the atmospheric surface layer above. The resistance analogue approach, it is possible to create quite complex, perhaps in the form of models of canopies (Shuttleworth, 1991), but the most complex models of soil-vegetation-atmosphere processes used in models of global climate rarely represent more than a layer of vegetation. They usually do so using equations which assume the behavior observed for the stomatal resistance of individual leaves, but "recalibrated" to describe the surface resistance of whole canopies (e.g. Sellers, 1989).

When rain falls on vegetation, the primary source of evaporating water is the foliage on the canopy, not the moist cells within the leaves as is the case for trees and vegetation. In effect, the stomatal resistance is "shorted out", and evaporation increases. The wetting by rainfall and subsequent drying of vegetation increases the resistance.

er the turbulent diffusion determined by wind speed ation and soil; and free ing the ground as sensible ily include simultaneous forced convection; and such energy fluxes without free e "first guess" estimate of

r atmospheric entities (e.g. oil-vegetation-atmosphere n equations are integrated differences in equivalent sensible heat), divided by a mplex models will include ough the boundary layer hey will include a "surface m inside to outside leaves; from within the vegetation layer above. Using the plex, perhaps multilayer, x models of soil-vegetation- y represent more than one which assume the general dual leaves, but they are anopies (e.g. Sellers *et al.*,

ting water is the free water is the case for transpiring out", and evaporation rates g of vegetation is usually

represented by a canopy water balance model, which includes the description of rainfall capture, of water dripping from the canopy, and of evaporation from the surface of the leaves. The most common form of mass balance used is that first described by Rutter (Rutter *et al.*, 1971; 1975), though most global models use a simplified version of this model.

Water movement within the soil is also described in models of soil-vegetation-atmosphere processes, albeit usually in simple terms, with the Darcy equation applied to compute vertical water flow between a few (usually 2 or 3) layers of soil. Furnishing values for the soil parameters required to enable a description based on the Darcy equation is arguably the most difficult problem in providing globally available, yet realistic models of soil-vegetation-atmosphere processes. Even at plot scale the variability in the value of these hydrological parameters from one place to the next is known to be very large indeed. The most common approach is to assign typical parameters by specifying the soil as belonging to one of (say) 10 classes. Very often the classification of Clapp and Hornberger (1978) is used, and small-scale and meso-scale heterogeneity in soil properties around these typical parameters is simply ignored.

In summary, models of soil-vegetation-atmosphere processes normally start by describing radiation exchange by treating this as two distinct streams of (solar and longwave) radiation. Having allowed for energy temporarily stored in the soil, the outstanding problem is to partition the remaining energy either as sensible heat, or for evaporation. The latent heat of vaporization is used to link the energy and mass balance equations, and diffusion models created to describe the movement of water vapor and sensible heat, these being implicitly or explicitly framed in terms of analogue resistances. Once energy and mass movement has been described for one time step in the model, the state variables (*i.e.* the stores of water and energy in the vegetation or soil) are updated, and these new values are then used to compute energy and mass flows at the subsequent time step.

Clearly the above general approach to describing soil-vegetation-atmosphere processes is capable of very diverse implementation, and there are numerous models of varying complexity currently available. Arguably the most important, unresolved issue is the

definition of which of the several processes outlined above must be represented, what level of accuracy. Further, if a process is necessarily in the model, how can parameters in the equations describing that process be specified globally?

3. Range of Complexity in Present Day Models

The broad range of complexity used in current models of soil-vegetation-atmosphere processes is in part generated by their diversity of purposes. Originally such models were mainly rather simple descriptions of evaporation for use in hydrological models. In the last decade their use in atmospheric models has expanded rapidly, both in magnitude and importance. In the future such models will increasingly be needed to provide vital coupling in models which include descriptions of both atmospheric and hydroecologic processes. Notwithstanding this diversity, as a basic minimum they describe

- bulk momentum exchange
- exchange of radiation (both solar and longwave)
- partition of radiant energy between soil heat flux, and latent and sensible heat flux.

The number of models of soil-vegetation-atmosphere processes described in the literature is large and expanding - there are at least 60 distinct models, perhaps 100, or more. Many are locally relevant "research" models with process emphasis relevant to specific research issues. The range of models applied at the global scale (in General Circulation Models) is less extensive, and they are usually one dimensional, mainly representing only a few soil and canopy layers. They exist in three general categories, namely

- (a) "Bucket" models, with no representation of a vegetation canopy;
- (b) "Micrometeorological" models, with substantial representation of canopy features; and
- (c) "Intermediate" models, with some canopy features included.

These three classes are overviewed in the following sections.

Bucket Models

The simplest models of evaporation were originally used in hydrological models and describe evaporation rate as potential evaporation moderated by soil moisture. Such models were also the first to be incorporated into atmospheric models (e.g. Manabe; 1969), with the addition of features and parameters relevant to energy and momentum, specifically with inclusion of

- albedo (to describe solar radiation capture)
- emissivity (to describe longwave radiation exchange)
- roughness length (to describe momentum exchange)
- soil thermal properties (to describe temporary storage of energy in the soil)

In such models energy partition is determined by calculating actual evaporation from potential evaporation, with sensible heat exchange then given as the residual in the energy balance. The moderating factor (in the range 0 - 1) used to scale actual to potential evaporation is set to be the fractional fill of a "bucket" of specified capacity, typically around 150 mm. This fractional fill is calculated from a running water balance between precipitation input and evaporation loss.

Micrometeorological Models

The most complex models currently used in General Circulation Models are typified by the Biosphere-Atmosphere Transfer Scheme (BATS) (Dickinson *et al.*, 1986) and the Simple Biosphere (SiB) Model (Sellers *et al.*, 1986). These are essentially micrometeorological models capable of representing surface exchanges reasonable realistically for a uniform plot of vegetation, but they are applied at a much bigger scale to areas up to (say) 10^5 km.

The exchange of solar radiation is represented in detail in BATS and SiB using the two-stream approximation (Ross and Monteith, 1975) mentioned earlier, while upward longwave radiation is calculated from the Stefan-Boltzmann law with a prescribed value of emissivity, and an effective radiometric surface temperature, this being an appropriately weighted combination of the surface temperatures of the vegetation and

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the soil surfaces in the model. Both models describe momentum transfer trade in terms of aerodynamic roughness which is prescribed for each cover type in BAT calculated during a pre-processing phase from prescribed canopy morphology in of the SiB model.

Transpiration is calculated in BATS and the SiB model, and in micrometeorological models, using a "big leaf" model, with whole-canopy resistance (effectively) represented with a Jarvis-Stewart model (e.g. Stewart, 19 product series of soil and vegetation cover dependent stress functions. micrometeorological models adopt a derivation (usually a simplification) of the model of canopy water balance to describe wet canopy evaporation. The parameter in describing this last process is the storage capacity of the canopy, normally defined as proportional to the leaf area of the vegetation, and tests again data (Sellers *et al.*, 1989) suggest use of values in the range 0.1 to 0.25 mm per area. Micrometeorological models include a simple, vertical soil hydrology, with balance calculation typically for 2 or 3 soil layers.

In general, micrometeorological models require specification of many parameters, different canopy covers and soil types, and accomplish this by allowing a large number of classifications of each. Never the less, many land cover classes (BAT has 12), and soil types (typically 10) are allowed. The primary problem with micrometeorological models is that the values of these parameters largely are speculative.

Intermediate Models

Intermediate models trade rigor in process description for simplicity. Common models allow a significant range of vegetation covers and soil types, but simplify physical representation of the exchange processes for each. The UK Meteorological Office model, for instance, prescribes fixed values not only for radiation and aerodynamic parameters for each land cover, but also for surface resistance (Lean and Viner, 1985). Soil moisture restriction is allowed for as a multiplicative function relating

entum transfer traditionally, each cover type in BATS, but opy morphology in the case

model, and indeed all with whole-canopy surface model (e.g. Stewart, 1988) as a nt stress functions. Most simplification) of the Rutter y evaporation. The critical apacity of the canopy: this is etation, and tests against field e 0.1 to 0.25 mm per unit leaf cal soil hydrology, with water

ition of many parameters for this by allowing a restricted d cover classes (BATS has 20; primary problem with the use parameters largely remain

or simplicity. Commonly such nd soil types, but simplify the ach. The UK Meteorological for radiation and aerodynamic ssistance (Lean and Warrilow, plicative function related to the.

moisture content within a rooting layer. Other models, such as that of the French Meteorological Service (Noilhan and Planton, 1989), are similar, but parameterize surface resistance in a way broadly equivalent to BATS.

An alternative approach is merely to simplify the computational processes involved in more complex models. "Simple SiB" (Xue *et al.*, 1991), for example, essentially uses the same assumptions as the SiB model, but codes these more efficiently by using empirical parameterizations to replace some of the time consuming direct calculations used in the original version of SiB. At the other end of the spectrum, bucket models can be made significantly more realistic simply by including a typical value of surface resistance [which is usually about 70 cm^{-1} , see Shuttleworth, 1993], to give a so-called "super bucket" model.

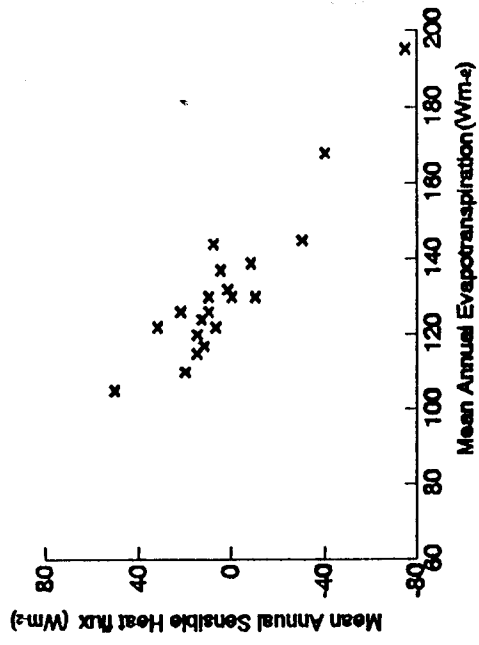
The use of this class of models with intermediate complexity is increasing, primarily because they represent a practical but still plausible alternative to more complex models pending evidence of the need for the more complex micrometeorological models.

4. Intercomparison of Present-Day Models

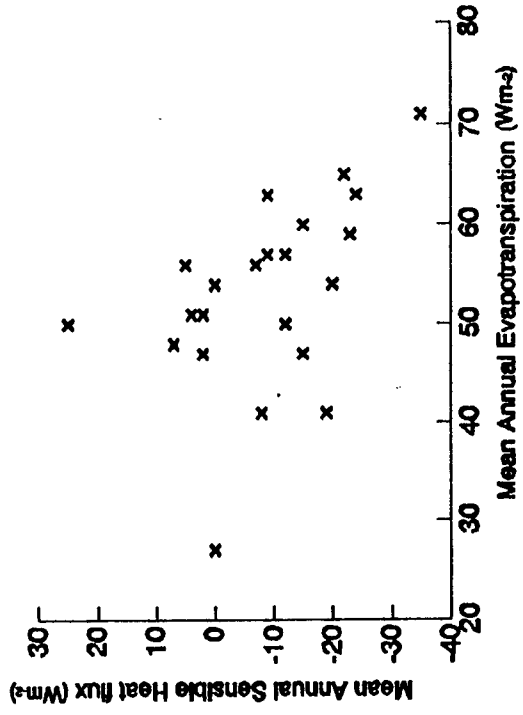
The Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS) is an ongoing international initiative to evaluate the performance of sub-models of soil-vegetation-atmosphere interaction processes in comparison with each other, and in due course in comparison with relevant field data (Henderson-Sellers and Dickinson, 1992). This project is still in an early stage of implementation, but has already provided interesting first results (Pitman *et al.*, 1993) which are summarized below.

PILPS currently involves around 25 land-surface parameterization schemes with a wide ranging range of complexity, and includes examples of the three classes overviewed in the previous section. So far these several models have been used with two synthetic series of climatological data calculated using a General Circulation Model, namely the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM). The two data series was selected to correspond to a temperate grassland site (at 52°N , 0°E), and a tropical forest site (3°N , 60°E).

(a)



(b)



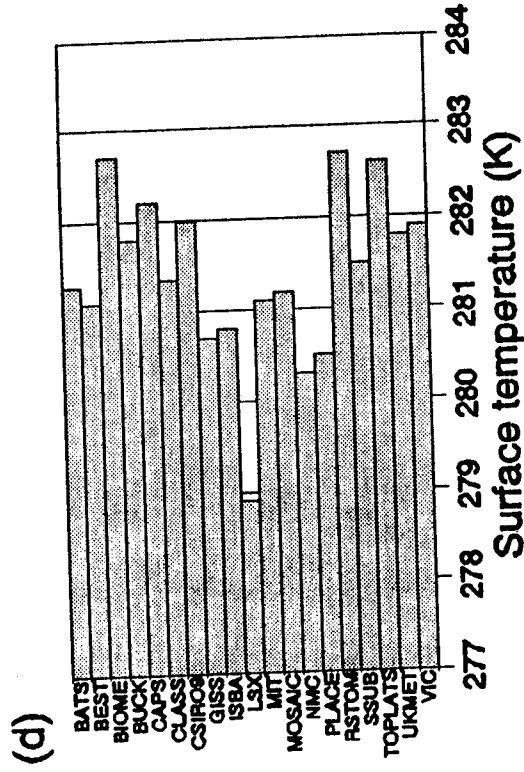
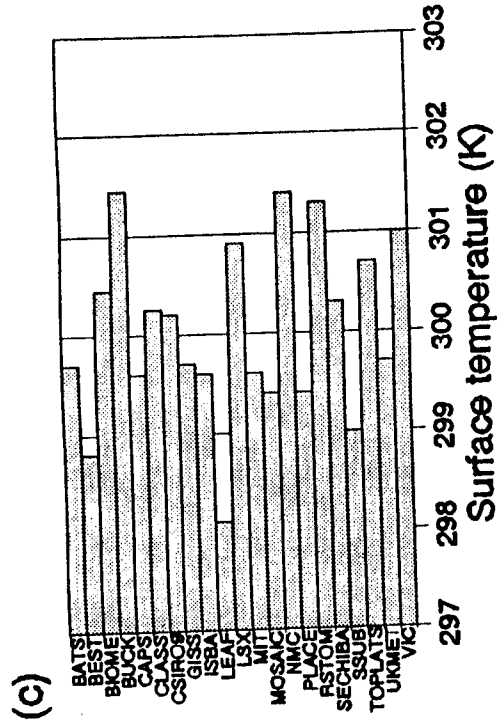


Figure 1. Mean annual evapotranspiration and sensible heat given by models involved in PILPS calculated from near surface weather variables for (a) "tropical forest", and (b) "grassland" sites. Mean annual temperature for (c) "tropical forest" and (d) "grassland" sites.

These data series were used as "driving variables" which were input to the seven surface parameterization schemes, and the resulting surface exchanges calculated. The outgoing fluxes from the ground are decoupled from the atmospheric case. That is, the driving variables remain those calculated by the CCM, unaltered by the calculated surface response. This will tend to emphasize distinction between calculated fluxes.

Figure 1(a) and 1(b) summarize the results of the PILPS comparison for the average values of latent and sensible heat for forest and grassland sites respectively. Figures 1(c) and 1(d) show similar results for mean annual temperature.

There are some very substantial outliers in these diagrams, and still substantial in the order $\pm 20 \text{ Wm}^{-2}$ for both forest and grassland amongst the remaining models if we neglect these outliers. The essential climatological difference in partition between forest and grassland is in fact resolved, but only just. Intercomparison of annual temperatures is not good - but neither is it. Neglecting outliers again, the agreement is to within about $1 - 2^\circ\text{K}$. However, it is important to remember that predictions of the global warming associated with an increase of atmospheric CO_2 concentration are themselves only in the order $2 - 4^\circ\text{K}$. In made similar comparisons between models at monthly and daily time scales. In the agreement, already marginal at the annual time scale, is reduced at shorter scales (Pitman *et al.*, 1993).

The difference in performance between land surface parameterization schemes in PILPS is not strongly correlated with the complexity of representation used in the models, except that some of the most significant outliers are bucket models. The origin of discrepancies is not known at this time, but the results of PILPS must raise the question as to which matters most, the complexity of process representation used in the models or the specification and calibration of the parameters they contain?

5. Calibration of Soil-Vegetation-Atmosphere Models for Simple Cover

In selected situations it is possible to define worthwhile studies of the potential consequences of land cover change when currently uniform biomes are at risk of large scale removal and replacement with alternative vegetation types. The potential removal of tropical forests for use as pastureland is an obvious example of dramatic (but not totally unrealistic) human-induced change, which has been much studied using General Circulation Models (e.g. Lean and Warrilow, 1989; Nobre *et al.*, 1991). Since it is change in the soil-vegetation-atmosphere interactions which is the essence of these studies, it is the calibration of models of these reactions which is arguably a most relevant issue in this case. Fortunately, this is one case in which provision of data for the calibration of models has been pursued with vigor (Shuttleworth *et al.*, 1991), at least in the case of the Amazon basin.

The Amazonian rainforest is still 90% intact, though deforestation proceeds quickly, mainly around the southern and western edges. Detailed micrometeorological measurements at a single study site in central Amazonia made more than a decade ago (Shuttleworth *et al.*, 1984; Shuttleworth, 1989) provided an at that time unique, set of data relevant to the calibration of models of soil-vegetation-atmosphere interactions. They included routinely available above-canopy weather variables measured on a scaffolding tower (equivalent to the "driving variables") for 25 months, together with directly measured surface radiation and latent and sensible heat exchanges for four extended (typically 2 months) periods selected to sample the seasonal precipitation regime. The routine measurements also included the net loss of rainfall resulting from interception by the forest canopy, and measurements of soil moisture status to a depth of 2 m, sampled at two weekly intervals.

Early comparisons between these data and then existing, uncalibrated land surface parameterization schemes suggested that these were in substantial error (Shuttleworth and Dickinson, 1989). But, subsequently these data were used to re-calibrate the parameters in one advanced scheme, SiB (Sellers *et al.*, 1989), and this substantially improved the descriptive ability of that model with respect to simpler, uncalibrated bucket models (Sato *et al.*, 1989).

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Although the above described experimental study provided useful calibration data for "before" in deforestation studies, data for calibrating the pastureland "after" studies has only recently become available (Shuttleworth *et al.*, 1991). Ongoing under the Anglo-Brazilian Amazonian Climate Observational Study (ABRACOS) providing sustained comparative measurements of near surface climate over extensive areas of undisturbed forest and cleared pastureland at three sites, which the ecoclimatological (precipitation) gradient across Amazonia. These measurements being complemented by comparative measurements of surface exchanges for forest and pasture during selected periods.

Although data from ABRACOS is only now being used to calibrate GCMs, some early results already provide categorical evidence of distinct differences in key parameters which are bound to have effect in deforestation studies. The difference in between forest and pastureland is quantified and confirmed, and this, together with characteristic difference in surface temperatures for the two vegetation types, provides a measurable difference in the net capture of radiant energy (Bastable *et al.*, 1993) because of the difference in aerodynamic interactions of forest and pasture land. Climate over pasture shows much greater daily variation than that over forest (*et al.*, 1993).

There is now known to be a very substantial difference in the rooting depth of forests and grassland, much larger than hitherto thought, with most of the root system water in the top meter under pasture, but forest able to access water to a depth of perhaps 6 - 8 meters. As a result Amazonian rainforest rarely shows evidence of stress in current climates, but the replacement grasslands respond noticeably to a change of a few weeks. In the course of the next few years the ABRACOS data will provide the credibility of presently available predictions of the potential consequences of large scale Amazonian deforestation, though not perhaps alter the predicted magnitude of likely change greatly. Currently the predicted regional changes are for about a reduction in evaporation, about a 30% reduction in precipitation, and about a 1°C increase in temperature.

6. Soil-Vegetation-Atmosphere Interactions in Weather Forecast Models

Until recently weather forecast models have assigned limited importance to their description of soil-vegetation-atmosphere processes, the reasoning being that forecasts are primarily determined by the data input at the beginning of a predictive cycle, and that surface inputs have little time to modify the ensuing predictions. This is largely true, and is particularly true for short term forecasts, but recent evidence suggests that the form of land surface parameterization schemes can be important since data to define the initial status of soil moisture stores is limited.

In general terms, it is the realism of the process representation which matters most in climate (*i.e.* long term) prediction. Faults in equations or parameter values are eventually revealed at "climate" time scales, but the initiation of moisture and energy stores at the beginning of climate prediction runs ultimately has limited effect. On the other hand, it is the initiation of such stores, particularly the initiation of the soil moisture store, which has most effect in weather prediction since this has impact immediately. In practice soil moisture status is not a data variable which is currently amenable to measurement, so the initiation of weather forecast models is derived from the predictive history of the model itself. In this way, gross defects in land surface parameterization schemes can propagate to calculate an erroneous initiation of the soil moisture store, which has an immediate and detrimental impact on the predicted weather.

A recent change in the formulation of land surface parameterization used at the European Centre for Medium Range Weather Forecasts illustrates this subtle effect well. Improvements in the realism of the land surface scheme between "Version 47" and "Version 48" of this model resulted in more realistic, wetter soils in the continental interiors during summer months. The resulting improved initiation of predictive runs brought dramatic improvements in forecasts of the 1993 floods in the Mississippi basin (Beijaars *et al.*, 1993).

Recent detailed studies of the model mechanism which allow better prediction (Viterbo, 1994) show that this is not due to greater evaporation and recycling of water locally, rather that the greater evaporation upwind (in Texas) allows a vertical temperature

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7. Investigation of Heterogeneous Vegetation Covers

Regardless of their complexity, most models of soil-vegetation-atmosphere processes virtually all land-surface parameterization used in GCMs represent surface exchange vertical flows, often assuming uniform land cover over extensive areas. In reality large portions of continents surfaces are heterogeneous, and this is particularly the case for land areas where human intervention is greatest. Heterogeneity occurs in overlying vegetation cover and in the underlying soil, and this is further complicated by topography, which can generate movement in surface and subsurface water, and which may in part determine the spatial position of vegetation.

In addition, atmospheric processes can themselves impose (albeit perhaps transient) surface heterogeneity by modifying surface radiation in response to cloud cover, and distribution of precipitation non-uniformly. Over the past decade most research on heterogeneous covers has focussed on investigating the representation of mixed vegetation, and, in particular, on determining the importance (or otherwise) of atmospheric advection in determining the values of area-average, effective parameters to be used in models of soil-vegetation-atmosphere processes.

One approach to representing heterogeneous surfaces is to decrease the intergrid distance used in representational models, but this rapidly generates huge demands on computational resources. The alternative approach is to seek to provide effective values for parameters within the existing land surface scheme such that these allow the model to describe area-average surface exchanges with adequate accuracy. This last approach is, in effect, a direct extension of the philosophy already used in providing plot-level calibration as described earlier. To pursue it requires that we postulate "aggregated rules", that is hypothetical and hopefully simple equations, which can be used at the scale to calculate effective area-average values of parameters relevant to the individual covers from the values present in the heterogeneous landscape.

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In practice the aggregation rules relevant to a particular land surface parameterization will tend to be scheme specific, but Shuttleworth (1991) suggests that a general approach to specifying the required rules might be to define the effective parameter as the weighted average for component covers using that function of the parameter which most nearly links it to a linear average of the associated flux. This suggestion reflects the fact that it is the area-average flux which is required.

When applying this general rule it is necessary to have some basic understanding of how each parameter within a given model changes the calculated flux. When defining this it is sometimes useful to frame the models as comprising components which directly relate to the availability of energy and water to the atmosphere. Using this approach, Shuttleworth (1991) postulates hypothetical aggregation rules for certain key parameters which control the capture of energy and water from the atmosphere such as albedo, emissivity and infiltration rate; for the magnitude of energy and moisture stores in the model; and for parameters which control the rate of access by the atmosphere to these stores, namely aerodynamic and stomatal resistance.

Much recent research has sought to investigate the validity or otherwise of the aggregation rules as described above. The methodology has been to combine experiment and representative models to synthesize area-average values of near-surface weather variables and surface exchanges, and then to compare these synthetic surface "data" with values derived using a single point-model in which parameters have been calculated using the hypothetical aggregation rule under test. Large scale field experiments (see Shuttleworth, 1991a, 1991b for reviews) have been carried out in which plot scale models of component covers in the selected study area are calibrated, and which allow the validation of meso-scale atmospheric models from aircraft and boundary layer sounding systems.

Analysis and interpretation of the first of these experiments, HAPEX-MOBILHY (André *et al.*, 1988) is most advanced. The primary land cover distinction in this case was between an extensive area of pine forest on the one hand, and an area of mixed agricultural crops on the other. Model simulations of the observed data has been very successful, and suggest that coupled surface-atmosphere models operating at meso-scale

with a grid mesh of about 10 km, can indeed adequately represent both the exchanges and ensuing atmospheric response well.

Similar large scale field experiments have been carried out elsewhere. The First Field Experiment (FIFE: Sellers *et al.*, 1988) investigated a 15 km x 15 km comparatively uniform prairie grass in Kansas, USA in 1987 and 1989. More field experiments have been carried out at the 10,000 km² scale in desert threatened areas with heterogeneous vegetation in central Spain, in 1992 (EFED *et al.*; 1993), and in the Sahel in 1993 (HAPEX-Sahel: Goutorbe *et al.*, 1994). and interpretation modeling of the data from these experiments is well in hand already contributing to define the need for a change in focus within this subject overviewed in the next section.

8. Current Status and Future Emphasis

Recently, the Interactional Geosphere-Biosphere Programme's Cure Project *Aspects of the Hydrological Cycle* and the *International Satellite Land Surface Climate Project* jointly sponsored a workshop, the *Tucson Aggregation Workshop*, whose was to provide a state of the art review of progress in the area of providing a description of mixed land cover. In practice this workshop provided better definition of the current status of our understanding of the following questions:

- a) is atmospheric advection important when formulating area descriptions of mixed cover?
- b) is topography important to area-average descriptions?
- c) can we remotely sense important aspects of vegetation cover exchanges?
- d) what are the primary future research priorities?

We overview the results below.

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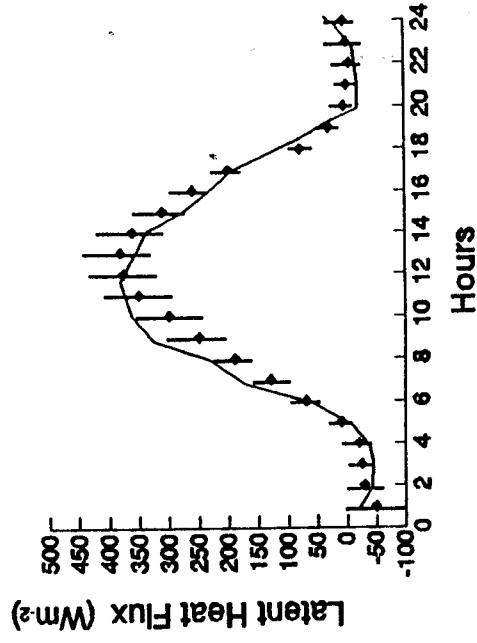
Is Atmospheric Advection Important?

Of the several question just posed, this is the question which has received most attention and for which there has been most progress. As outlined above, the method used to address this question is to use coupled surface atmosphere models to synthesize "true" area-average fluxes, and to compare these values with those calculated from a single equivalent point model, whose parameters have been defined using hypothetical aggregation rules. Experiments provide reassurance that the coupled model is a reasonable representation of reality. Based on observations made during HAPEX-MOBILHY, which suggest that coherence in surface cover can effect atmospheric processes (Shuttleworth, 1988), researchers have been careful to implement this methodology using models which represent atmospheric response at two different spatial scales, *i.e.* mesoscale circulations at greater than about 10 km, and plot scale movement at much finer scale.

Noilhan (1994) made a model synthesis of the data from HAPEX-MOBILHY using a meso-scale model with a grid mesh of 10 km, this being consistent with the length scale of the main organizational structure is land cover in this experiment. The model was parameterized using soil/vegetation maps for the study area, some of the information in these maps, particularly that relating to vegetation, being derived from satellite data. Figure 2 shows that the synthetic area - average fluxes calculated with this representational model agree with those calculated from a single point model with aggregation rule derived parameters to within about 10%, both in terms of mean daily trend and in terms of the integrated evaporation over several weeks. The implication is that mesoscale atmospheric advection only has limited impact at this level, even for the dramatic land cover distinction between forest and agricultural crops, and where substantial mesoscale structure is observed (and modeled) in the atmosphere.

Similar studies have been made using coupled surface-atmosphere models to investigate the influence of atmospheric advection at patch scale, that is with vegetation patches in the order of a kilometer, or less. In fact valuable initial insight had already been gained using coupled models operating at this scale with idealized and/or extreme mixes of surface covers. Mason and co-workers (Mason, 1988; Woods and Mason, 1991) suggested

(a)



(b)

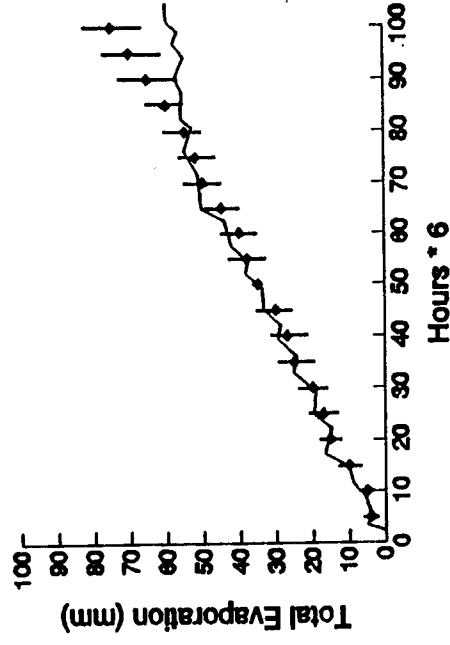


Figure 2. (a) Latent-heat-flux for the HAPEX-MOBILHY site calculated using surface scheme with "aggregate" parameters (full line) compared with the mean and deviation for grid squares in a meso-scale meteorological model; (b) integrated evaporation for the HAPEX MOBILHY site given with "aggregate" parameters (full line) compared with measurements for sample sites in the study area.

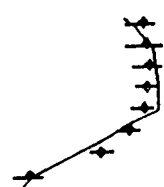
that an aggregation rule based on the square of logarithms is relevant for calculating area-average values of aerodynamic roughness. Blyth and co-workers (Blyth *et al.*, 1993), using a development of Mason's model which incorporated a simple Penman-Monteith description of surface energy balance, suggest an aggregation rule which assumes parallel summation of surface resistance was adequate to 5 - 10 % for reasonable mixtures of dry vegetation, but that more complex schemes may be needed if surface water (from rainfall) is present in small patches within the landscape.

More recently, Arain (1994) has investigated patch scale aggregation rules relevant to the BATS model for plausible mixes of vegetation relevant to the FIFE experimental site. The philosophy believed his approach is that, in practice, ecoclimatological limitations mean that the range of cover types relevant to real life, real world situations is likely to be a restricted subset of all those possible within a particular land surface parameterization scheme. In general, the results of this study confirm that simple aggregation rules work extremely well to describe area-average behavior even with an advanced land surface parameterization such as BATS. But studies with artificially wetted (irrigated) patches in an otherwise dry landscape reveal an interesting exception. In this case atmospheric advection remains a small effect, but the non-linearity of the relationship between soil moisture and plant wilting point in BATS means that the true area-average energy balance actually alters in a non-linear way in a landscape with extreme differences in soil moisture.

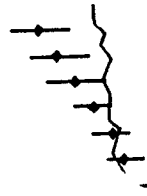
This last result points the way for a future emphasis in research attention, away from the current focus on atmospheric advection processes, towards a focus on subsurface variability in accessible soil moisture in the rooting zone, and towards possible topography-related movements in this.

Is Topography Important?

Research into the importance of topography in the formulation of area-average soil-vegetation-atmosphere processes is much less advanced than that into the importance of vegetation cover. However important initial progress has been made in understanding the effect of topography-induced changes in near-surface weather variables.



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Remote sensing data can be used in two main ways. It can be used in its now we
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Current evidence suggest that if we are to use remote sensing data to monitor paramet
relevant to soil-vegetation-atmosphere relationships, we will need to exploit both satell
mapping capability and calibrated algorithms simultaneously. To take the example
albedo, satellites can measure reflected radiant energy over the wavelength relevant
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require at least some knowledge of the vegetation class prevalent there.

In a similar way information on the fractional cover of green vegetation derived from
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type and structure of the vegetation, and by the spectral properties of the underlying s
Again, any algorithms derived to provide fractional vegetation cover from satellite d
are likely to be sensitive to view angle, and so to be cover class dependent to some deg

Raupach and co-workers (Raupach, 1994) made modeling studies in which they used coupled surface-atmosphere model to synthesize area-average exchanges in a landscape of moderate (about 100 m) topography, but with surface cover and soil water availability was prescribed as constant across the landscape. The model therefore allowed for topography generated changes in the surface radiation balance, and in near-surface temperature, humidity and wind speed. The results of these studies are valuable in that it demonstrates that area-average exchanges are little effected by spatial variability in these meteorological variables alone.

Notwithstanding this last simple result, very substantial unresolved questions remain regarding topography which are high on the agenda of the global research community for the coming decade. Specifically:

- a) what are the consequences of topography - generated differences in soil moisture on area-average exchanges?;
- b) what are the consequences of topography - generated differences in vegetation cover?;
- c) do the topography-generated differences in soil moisture and vegetation, and those in meteorological variables interfere with each other?; and
- d) how do all of the above processes alter in a region with marked (as oppose to moderate) topography.

Can We Remotely Sense Aspects of Soil-Vegetation-Atmosphere Relation Independently of Cover Class?

Certain aspects of soil-vegetation-atmosphere relations may be amenable to indirect measurement from satellite systems. Most obvious amongst these parameters are the area-average albedo, the fractional vegetation cover and the level of photosynthetic activity. Clearly there are substantial problems of a technical nature to be resolved before these can be measured - related to the calibration of sensors and sign

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established mapping mode, to define the location and areal extent of different land
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these last relationships universal and independent of cover class?: some years ago remote
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Current evidence suggests that if we are to use remote sensing data to monitor parameters
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type and structure of the vegetation, and by the spectral properties of the underlying soil.
Again, any algorithms derived to provide fractional vegetation cover from satellite data
are likely to be sensitive to view angle, and so to be cover class dependent to some degree.

Finally, there is evidence that the net absorption of photosynthetic radiation by plant stands provides some information on the surface resistance of the stand (see Sellers *et al.*, 1992). However, this relationship is strongly determined by basic physiology used by the plant to effect photosynthesis. Figure 3 illustrates an example for the FIFE site, where the relative proportion of C_3 to C_4 plants at a particular prairie grass site substantially alters the relationship between an important radiation control on surface resistance, and the "simple ratio" between reflectances in the near infrared and visible regions of the spectrum.

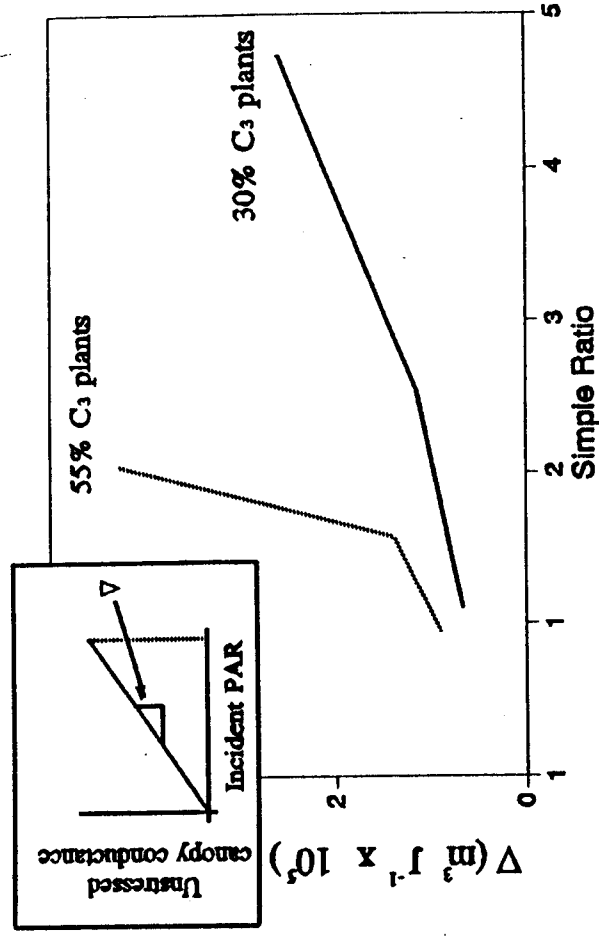
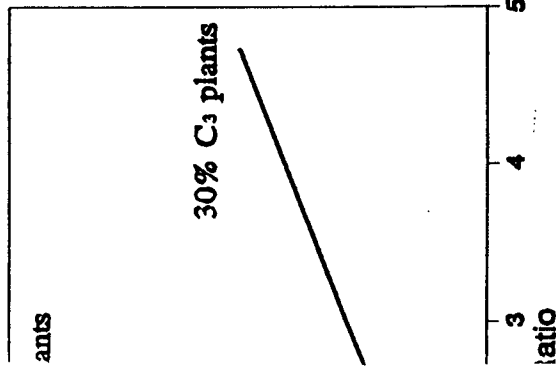


Figure 3. The sensitivity of unstressed canopy conductance to incident Photosynthetically Active Radiation (PAR) observed at two FIFE sites with different mixes of grass cover in relation to the measured simple ratio at those two sites.

Future Research Priorities?

On the basis of discussion at the Tucson Aggregation Workshop, what is the required shift in research emphasis?

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variance to incident Photosynthetically
 with different mixes of grass cover in

Clearly there is a need to refocus studies of heterogeneity away from atmospheric advection processes, towards better defining the relation between vadose zone hydrology and biospheric processes, and towards the influence of subterranean variability on average interactions. Similarly, there is need for some shift in observational emphasis away from the current preoccupation with short-term sampling in the spatial domain towards longer term periods of data collection, so as to better measure and understand seasonal and inter-annual variations for important vegetation cover classes.

Studies of the effect of topography are needed, and experimental and modeling studies in regions with marked topography, and studies to investigate the interaction between topography-related changes in meteorology and those in vegetation and soil moisture are a high priority.

In the area of remote sensing, we need to recognize and use the general result that it is necessary to simultaneously exploit both the mapping capability of satellites, and the capability to provide globally available fields of surrogate measures of parameters important in soil-vegetation-atmosphere relations.

Workshop, what is the required shift

Hydrological Models, Regional Evaporation, and Remote Sensing: Let's Start Simple and Maintain Perspective

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1. Introduction

This paper seeks to be provocative in the sense that it is intended to stimulate a line of thought and perhaps a climate of opinion in which the global modeling of terrestrial interactions, evaporation in particular, is predisposed towards simplicity. The justification for adopting this intention is an opinion that, having been faced with the new challenge to provide an acceptably realistic description of hydrological process at a global scale, but having simultaneously been faced with powerful visual impact of remote sensing images to remind them of the heterogeneity of terrestrial surfaces, hydrologists may have suffered some loss of confidence and an associated lack of perspective on the scale of the problem to be addressed. Perhaps this loss of confidence is reflected in the phrase "The Trials and Tribulations of Modeling and Measuring" which is part of the title of this workshop.

In addressing our new and exciting challenge, we should draw courage and motivation from the fact that hydrological models are already routinely applied with considerable success to practical problems at catchment scale. Often the point-to-point variability in soil characteristics across an ostensibly uniform 'patch' of vegetation or hill slope is huge, and the plot-to-plot variability in vegetation cover across a catchment can itself be substantial. Yet hydrologists continue to triumphantly build and apply models to catchments. In fact, much of the land surface heterogeneity which gives concern is already present at subcatchment scale, whether visible in remotely sensed images or otherwise.

Within scientific hydrology, as we have acquired more detailed understanding of hydrological processes over the last 25 years and as computer resources became more

available, our first instinct was to seek to build more and more complex models. With time, we learned that our ability to provide proper initiation and calibration of the parameters used in such models was a severe and fundamental limitation. We learned to strike a balance between detail in model description and the justifiable data input we could make to them. Shouldn't we transfer that acquired common sense and experience when approaching this new challenge? Here I hope to speed progress towards reaching the equivalent compromise when hydrologists address global change by providing a positive counterbalance to any overestimation of the scale of the challenge we face in providing adequate hydrological models and in providing representation of the evaporation process in particular.

This I do, first, in Section 2 by distinguishing several distinct objectives behind seeking improved representations of hydrological processes in the context of better modeling Global Environmental Change (GEC). The point is that, if there are distinct objectives, then there may well be distinction in the complexity of the modeling required to achieve each objective. Explicit representation of slopes and surface features is presumably necessary in the hydrological models which give interpretation of predicted global change in terms of modified surface and subsurface flows, but these models can be locally specific. On the other hand, feedbacks and mixing in a dynamic, fluid atmosphere might mean that there is little need for *explicit, geographically accurate* description of aspects of land surface heterogeneity within atmospheric General Circulation Models (GCMs) to improve climate change prediction. For instance, there is now substantial evidence that heterogeneity in land cover, providing it has appropriate size, can generate enhanced vertical transport in the atmosphere by stimulating mesoscale circulations (e.g., Dalu and Pielke, 1993; Avissar and Chen, 1993). In practice, perhaps, this enhancement can be parameterized by defining semi-empirical functions involving the statistics of land cover distribution (Zeng and Pielke, 1995). Equally, the complexity required when describing area-average surface exchanges in GCMs is arguably less than that required for interpretative hydrological modeling, because land surface hydrology is just one of several influences on climate, and GCMs must retain an appropriate balance in the way they include these influences. In this case, an appropriately defined averaging procedure applied to component land covers in each grid square might suffice to define the

effective values of relevant land surface parameters (Shuttleworth, 1991, 1993; Noilhan and Lacarrere, 1994).

Over the last 5-10 years, there has been an explosive proliferation in the range of submodels [Soil-Vegetation-Atmosphere Transfer Schemes (SVATS)] which seek to provide a description of the energy-water interactions of terrestrial surfaces for the purposes of global environmental change prediction. There has also been a gradual but ongoing increase in the complexity with which they represent the physical and physiological processes involved. This proliferation gives more insight into human ingenuity and the now ready availability of computers than it does into the complexity which is truly required in SVATS. Just how broad a range of distinguishable land covers do we really need? What characteristics of these land covers really matter? We will look at these issues in Section 3 with an emphasis on seeking simplicity and on seeking greater relevance to satellite data, assuming that such data are a likely mechanism to affect global coverage of model parameters. Having this emphasis, we then consider recent experimental and modeling evidence which suggests that perhaps quite simple averaging schemes might provide acceptable aggregate representation of heterogeneous vegetation cover.

Part of the pressure from hydrologists to include hydrological complexity in GCMs (beyond that justified to improve climate prediction) is generated by the parallel need to allow more ready and more transparent hydrological interpretation of any changes they predict. The final part of this paper seeks to address this problem by advocating a computational mechanism which will allow GCM-generated, globally defined, vector fields of atmospheric humidity to be expressed in terms of the net input of water to irregularly shaped catchments. In this way, the persistently distinctive aspects of meteorological and hydrological models, namely their differing requirements for representational complexity and their optimum modeling elements (rectangular on the one hand, irregular topographically determined catchments on the other) can be reconciled, leaving each type of model free to develop in its own optimum and balanced way.

2. The Objectives of Hydrometeorological Modeling

Traditionally, the term hydrological models is used to describe a combination of mathematical formulae which provide a description of the movement of surface and subsurface waters within a catchment, from information on the precipitation input and other information which allows estimation of the evaporative loss from the catchment. Therefore, such models necessarily describe both horizontal and vertical water movement, with at least as much emphasis placed on modeling the former as the latter.

Until recently, the 'hydrological models' used in GCMs have been profoundly different to this. Their main objective has been to describe the vertical movement of that portion of the precipitation which is returned to the atmosphere by evaporation from the surface of leaves and soil or by transpiration through plants from the shallow layer of soil accessible to the roots. Here we broaden this narrow definition of 'GCM hydrology' to include the representation of both horizontal and vertical flows and, with this re-definition, there are at least four main purposes for carrying out hydrometeorological modeling, as follows:

- (1) To allow predictions of global environmental change to be better interpreted in terms of its impact on hydrological and agrohydrological systems.
- (2) To allow validation of the description of hydrology provided by General Circulation Models in present-day climates by making comparisons against historical hydrological records of rainfall and runoff.
- (3) To enhance the accuracy and credibility of predictions of global environmental change made with global models by improving the description of the land surface energy and water exchanges.
- (4) To enhance the potential for improved prediction of global environmental change by allowing better description of other surface exchanges controlled or significantly influenced by hydrological flow or soil moisture.

The following subsections, 2.1 to 2.4, separately discuss modeling needs in relation to the above objectives, along with opinions on the activity required to address them with *minimum complexity*. The alternative approach to satisfying these objectives is to seek to write hydrological components into models of global environmental change which address these four needs simultaneously. This might be assumed to be the best and most complete approach. However, it requires modeling with maximum complexity and modeling which rapidly becomes intractable (or at best inefficient) and which requires that large amounts of calibration data are available worldwide. A consistent argument in this paper is that the above distinct objectives are best considered separately and that, in each case, the approach should be to address the needs with a predisposition to start with the simplest assumptions consistent with physical and biophysical insight.

2.1 Hydrometeorological Models and Improved Interpretation

To allow interpretation of global change predictions in terms of hydrological and agrohydrological impacts, the primary need is for an interfacing procedure which can interpret the global fields of weather variables provided by GCMs in terms of the weather variables required by traditional hydrological models. These hydrological models can then operate independently in the local region of interest. Indeed, they may well already exist and may even have a relevant calibration based on existing data if the predicted climate change lies within the range of past climate variability.

Section 4 outlines a novel approach to providing the required interface between GCMs and local hydrological models. However, there are alternative approaches, sometimes called 'Weather Generators' (IGBP, 1993), although these tend to be more complicated and more location specific. They include the possible use of mesoscale meteorological models, perhaps operating off-line from a GCM at less than GCM grid scale, but driven by GCM-gridded data. Alternatively, they may access existing weather data, if available, to provide the statistical linkage between mean and distributed weather variables for the region of interest, under the assumption that such statistical linkages will not change substantially when the climate alters. However, the key point made here is that *none*,

of these potential linkage procedures requires that the hydrological models used for interpretation are directly coupled to the GCM providing the source data, thereby removing the need to compromise the operational efficiency of the GCM just to facilitate the hydrological interpretation of its output.

2.2 Hydrometeorological Models and Improved Validation

Existing hydrological records are never in an appropriate form for direct validation of the simplified hydrological schemes in GCMs because there is a basic inconsistency between both the spatial and temporal characteristics of the hydrological data on the one hand and the computed GCM outputs on the other. The easiest way to deal with this problem is to write some form of hydrological model which can provide credible routing of the GCM runoff computed for individual grid points through simplified river systems (e.g., Liston, 1993). In this way, derived runoff data can be synthesized which has appropriately delayed river flow timing for the few large rivers that can be represented at the grid mesh scale of GCMs.

It is clearly necessary that the representation of river routing is sufficiently credible and well calibrated that it is, in fact, the GCM--rather than the routing scheme--which is validated. Further, bearing in mind the probability that the grid mesh length used in GCMs will reduce in size over the next decades as computer technology advances, it is sensible to seek a robust, grid mesh size independent procedure to define the form of the river system represented and to calibrate the constants used to calculate the river flow against available hydrological records. Such a procedure will likely now be based on Geographical Information System (GIS) techniques (Moore et al., 1992). In the context of this paper, the relevant point is that, *for the purposes of validating a GCM against runoff data, the required hydrological modeling is essentially simple and large scale* and that there is no real need to couple even this simple routing model into the GCM. There may, however, be other reasons for doing this coupling, as described in Sections 2.3 and 2.4.

A further mode of validation might exist in which an advanced hydrological model is used to synthesize the area average values of some of the model-calculated variables which the hydrological model shares with the GCM. Such variables might include soil moisture, evaporation, or perhaps surface temperature. In this case, the primary requirement is that the model is realistic and credible, which might well mean that it is itself validated at sample points against patch-scale data and that it is used as an interpolation mechanism. To be realistic, the hydrological model used for interpolation may need to be complex but, as in the case of validation, it can operate independently of the GCM in the local region where validation is to be made. In this case, the hydrological modeling methods are not necessarily new--they are merely being applied over a somewhat bigger area than previously--and the most exacting novel requirement is not the model method; rather it is for distributed near-surface weather data with which to drive the model across the now larger validation area.

2.3 Hydrometeorological Modeling and Improved Prediction

Most of the considerable effort which has so far been directed towards improving the representation of hydrology in GCMs has focused on providing a more realistic description of the vertical movement of water. This is because it is these vertical flows which are most important in determining the net return and the rate of return of water to the atmosphere by evaporation. However, there has already been substantial activity in this subject area and there are, in the author's opinion, too many alternative SVATS, some with arguably too much complexity. However, we postpone further consideration of this last point until Section 3 and here focus attention on the as yet unfilled need to include at least some simple representation of horizontal water flows in GCMs.

Accessibility is the key word when addressing the need for an adequate representation of hydrological processes in atmospheric models (see Shuttleworth, 1991 and Section 3). It is the extent to which water is made accessible to the atmosphere and the ease with which it is accessible that are responsible for the sensitivity of weather and climate predictions to the manner in which hydrology is described in GCMs. Therefore, it is

only those aspects of horizontal flow processes that change the atmosphere's access to surface and surface waters which need to be included (along with existing SVATS) to improve the predictive ability of GCMs.

2.3.1 Modeling the Consequences of Horizontal Flows in GCMs

Within an individual grid square, water is accessible to the atmosphere only if it is on the surface of vegetation or soil, in moving or stagnant bodies of water, or in the near surface soil to depths which can be reached by the roots of actively transpiring vegetation. Once water drains from the rooting zone or enters the body of a deep river, it is no longer available for evaporation. However, it is possible for soil water lost to the rooting zone at the top of slopes to become accessible to the atmosphere again if hillslope-induced flow gives concentration of water lower down, so that water once again rises into the plants' rooting zone (see Figure 1).

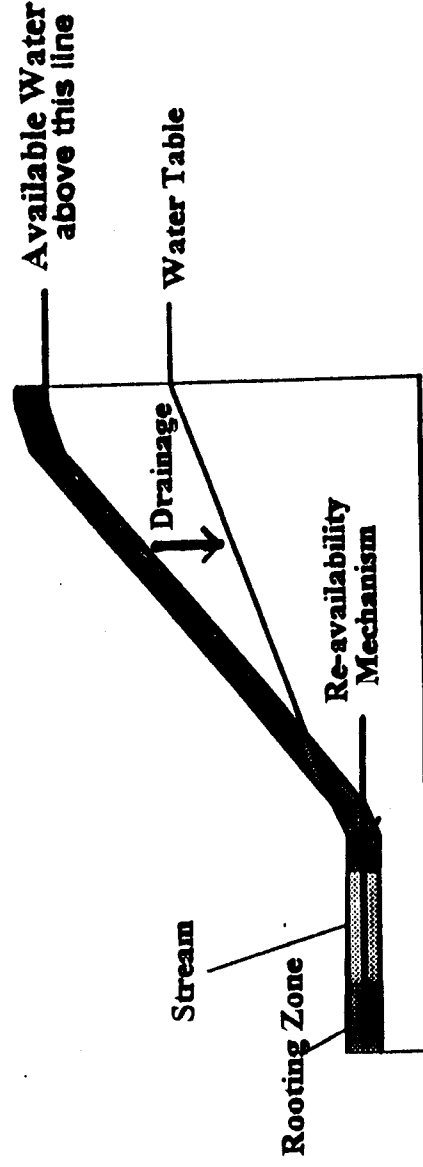


Figure 1. Schematic diagram illustrating the surface water 'available' for return to the atmosphere and the effect of horizontal, subterranean flow in making water which has drained from the plants' rooting zone at the top of a slope 're-available' at the bottom of the slope by returning it to the rooting zone.

Similarly, water which has been 'lost' to rivers or lakes can in time become accessible to the atmosphere again if the level of the water body drops by evaporation to reveal successively deeper layers of water. This process could be important, for instance, in the

case of many shallow, stagnant lakes in high latitude North America. Otherwise, stream water drains through the bed of the river or lake or it runs off down river out of the grid square and, for all practical purposes, it is no longer available to support evaporation from the grid square in question.

In principle, it is possible for water lost as river runoff or by deep drainage in a particular grid area to become re-accessible to the atmosphere elsewhere in the same grid square, if the rates of surface or subsurface flow and the underlying topography of the streams or aquifers are appropriately combined to give flooding or springs. In practice, the area of the free water surface so generated is unlikely to be more than a few percent of that of the grid square; this phenomenon can probably be neglected from the standpoint of improving GCM predictions. On the other hand, exactly the same phenomenon might be more significant to regional weather in the case of very large rivers, such as the Amazon and the Nile, which are prone to extensive seasonal flooding and which are sufficiently large to be acknowledged in the routing algorithms described in Section 2.2. In this case, water which has become inaccessible to the atmosphere in one grid area might become re-accessible again elsewhere in another grid area (see Figure 2).

From the restricted standpoint of providing better climate change prediction, it could be argued on the basis of the above discussion that the most important, and perhaps the only worthwhile, horizontal water flow related improvements which could be made in GCMs are the following:

- (a) In the case of grid squares with a marked topography, description of the topography-generated, intragrid, horizontal flow mechanisms which allow water that has drained from the rooting zone at the top of hill slopes to become re-accessible to the atmosphere again at the bottom of hill slopes, as illustrated in Figure 1.

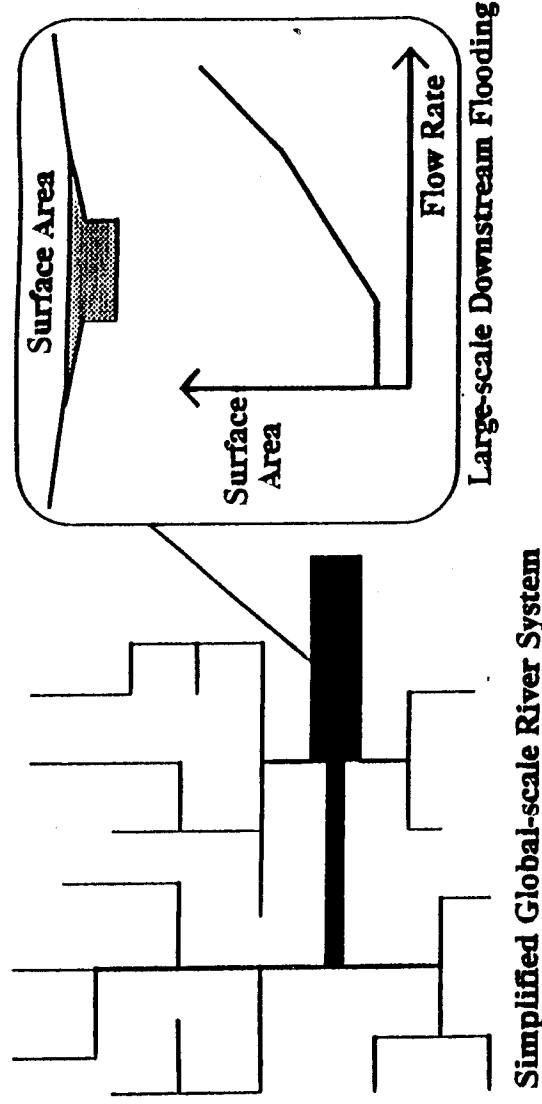


Figure 2. Schematic diagram showing a simple river routing scheme appropriate for incorporation in a GCM and how water which has been lost to the atmosphere by becoming stream flow in one grid square might become 're-available' to the atmosphere if it results in large-scale seasonal flooding in grid areas downstream in the simplified river system.

- (b) In the case of very large, flood-prone rivers, description of the intergrid flow mechanisms which allow water lost to river flow in one grid area to become re-accessible to the atmosphere in other grid areas because the surface area of the river is substantially increased as it spreads across the flood plain, as illustrated in

Figure 2.

In line with the philosophy of this paper, we should seek to accommodate description of the above two processes with minimum complexity. In this respect, the second, intergrid process is easier than the first. It merely requires the real-time coupling into the GCM of the simplified river systems discussed in Section 2.2 and then including the contribution of a river flow dependent, free water source area to the (vertical) return of water vapor to the atmosphere. Remote sensing measurements of river area and conventional measurements of river flow for the few relevant large rivers might provide calibration for this model feature.

Providing description of the first process, the intragrid process (a) above, is more difficult, but some useful, relevant work has already been done using fairly simple distributed hydrological models, particular TOPMODEL (Beven and Kirkby, 1979). The work of Beven and Wood (1983), Band and Wood (1988), Wood et al., (1990) and most relevant, that of Famiglietti and Wood (1994a) is valuable in this regard. Such research is starting to show that intragrid horizontal flows may impact the calculated area-average evaporation rate, at least in certain conditions (Famiglietti and Wood, 1994b), even though the area of the downslope wetted zones may, in practice, be only about 10% of the total surface area.

Such distributed hydrological models still retain a level of representational complexity for horizontal flows which, were it suggested they be explicitly incorporated into GCMs, might be considered inconsistent with the relative importance to global change prediction of this within grid, horizontal flow generated re-accessibility of water. This is yet to be determined. Could such models be used to define the effective values of a few topography-determined parameters in a simpler modeling approach which involves including a few 'representative slopes', perhaps one for each allowed vegetation cover, for which hillslope flow is modeled in each GCM grid square? Progress in this branch of research is rapid, and it is likely these questions will be answered in the next few years.

In any case, the suggestion made here again is that, for the purposes of improved climate and weather prediction, *the additional modeling in GCMs to acknowledge the relevant impacts of horizontal flow on the atmospheric return of water and energy should, and probably could, be simple. Simple description should at least be attempted first.*

2.4 Hydrometeorological Modeling and Other Surface Exchanges

The hydrometeorological modeling of river flows and soil moisture is a necessary component of the modeling of certain other surface atmosphere exchanges which are of significance in improving the prediction of global environmental change. The most

significant hydrology-related processes in this regard are perhaps changes in ocean salinity and the exchange with the atmosphere of biogeochemicals.

Accommodating the impact of (large) river flows on ocean salinity should be comparatively simple and is already being addressed in so-called "Climate System Models". The need is to incorporate the simplified river routing schemes discussed in Section 2.2 into the GCM and then to formulate an algorithm to describe the mixing of salt and fresh waters in those grid squares which subtend the river mouth of the world's major rivers.

The carbon cycle is arguably the most important global cycle after those of energy and water; providing a worthwhile description of carbon dioxide exchange over continental surfaces is currently a high priority (IGBP, 1993). It is extremely helpful that the release of water from plants (as transpiration) and the photosynthetic absorption of carbon dioxide are both strongly and jointly controlled by the same stomatal mechanism. It is also helpful that some aspects of this stomatal control seem to be related to measurements which can be made from space (see Hall et al., 1992; Sellers et al., 1992; and Section 3). In the context of our present discussion, the relevant point to make is that, because these important, largely vegetation-determined exchanges share a common dominant controlling mechanism, it is possible (indeed essential) that the models used to provide their global scale description should share a common level of complexity. If, as we argue in Section 3, the SVATS used to describe the energy-water exchanges of vegetation stands are to remain simple, then those used to describe large-scale carbon dioxide exchanges should be similarly simple.

Providing adequate hydrometeorological models to serve as components in the description of other biogeochemical releases will be altogether a more complex matter. This is primarily because some of the processes involved in such release are not likely to be linear functions of soil water content and because a significant proportion of those biogeochemical releases with most global significance, such as that of methane from swamps or rice paddies, may occur from areas so small that they are, in global terms, essentially point sources. These two factors argue for the use of detailed high-resolution hydrological modeling to address this issue, but it is not clear that implementing models

with this complexity across the globe is justified, especially if it were also proposed to include such a submodel in (GCM) real-time. Perhaps some compromise will emerge in the course of the next decade involving detailed submodeling for key source areas, but using a model time step which is less than that used in the atmospheric component.

In the specific case of biogeochemical release then, it is still too early to advocate a predisposition towards a simple modeling approach as for other processes; indeed, the best first step forward may be to write complex locally specific models. In doing this, it is essential not to lose sight of the true numerical significance of the calculated biogeochemical releases when viewed from the perspective of global balances. This is easily done.

In summary, when modeling changes in ocean salinity and net carbon photosynthesis at global scale, the recommendation is again to use models with simplicity consistent with that used elsewhere in predictive global scale models and which is consistent with the relative sensitivity of prediction to these processes. For other, essentially point source biogeochemical exchanges, the current recommendation is to "act local, but think global".

3. The Complexity of Soil Vegetation Atmosphere Transfer Schemes

Earlier, the opinion was expressed that there are now too many SVATS and that they are probably too complicated. Here I return to this point by again asking what about such submodels is most important from the point of view of modeling the surface-atmosphere exchange of energy and water vapor, this being currently the primary purpose of SVATS. In pursuing this argument, I draw heavily on ideas and opinions previously expressed in Shuttleworth (1991) and at an earlier NATO Advanced Study Institute (Shuttleworth, 1993).

3.1 What Matters?

It is first necessary to consider which aspects of the energy-water interactions of vegetation-covered continental surfaces are truly essential to include in SVAT submodels for the purposes of adequate climate and weather prediction. As stated earlier, the key characteristic of such aspects (and the parameters which determine them) is that they must strongly influence the extent to which energy and water is available to the atmosphere. Again, the key word is *availability*.

To become available, energy and water must be first captured at the land surface. The area-average value of albedo must be important, because this directly influences the amount of short-wavelength radiant energy captured at the ground that is available for return to the atmosphere. It is well known that land cover type can alter the value of albedo significantly. The surface emissivity is also important because this directly determines long-wavelength radiation exchange, but defining its effective, area-average value is less difficult than for albedo because the variation from one land cover type to the next is less. Clearly, the area-average maximum rate at which water can infiltrate the soil is important too, because water which runs off into rivers and lakes is generally more concentrated and less available to the atmosphere, as discussed in Section 2.3.

To remain available once captured, energy and water must remain in stores near the surface which are accessible to the air above, and the area-average magnitude of such accessible stores are clearly key parameters because they determine the net amount of energy and water available for return. The atmosphere is inherently a 'fast' system: it can respond to changes in surface inputs at the daily time scale--witness the daily cycle of convective rain. However, it also responds to slower changes if there is 'memory' in the stores of energy and water to which it has access. Therefore, it seems appropriate to acknowledge and include two types of stores in models of the land-atmosphere interaction. The first is the energy and water stored in comparatively easily accessible form in, or in the case of water on, the vegetation and soil surfaces. The second store is the energy and water held in less readily accessible form in the soil, but only to the depth where they are still (ultimately) available for return to the atmosphere.

Because the timing of the release of available energy and water is important, the control exerted by the vegetation and soil in determining the rate of atmospheric access to the stores of available energy and water is also important. Commonly, the mechanisms involved in this control are expressed in terms of a resistance network, which includes at least two resistances, one in the atmosphere and one associated with the surface (the latter controlling the rate of diffusion of water vapor through the stomatal apertures from inside plant leaves into the canopy air stream outside).

In summary then, the primary land surface characteristics which influence the magnitude and rate of access of the atmosphere to energy and water available at the ground are the albedo, a ; the surface emissivity, e ; the maximum surface infiltration rate, I ; the aerodynamic roughness, z_0 , because this determines the efficiency of the return of energy and water near the ground; the surface resistance, or its reciprocal the surface conductance, g_s , because this determines the efficiency of the return of soil water to the air adjacent to the plant leaves or bare ground; and the size of two types of stores for energy and water, one in the vegetation which is easily accessed, one less accessible in the soil.

Most of the many SVATS now available for use in GCMs include calculations which involve these key characteristics, although some (Shuttleworth, 1992) still follow the approach of many practicing hydrologists in not including explicit description of the stomatal control and rainfall interception mechanisms when calculating evaporation rate. It is not yet clear what level of mathematical complexity in SVATS is justified by an associated improvement in the predictive ability of the GCMs in which they reside. It is likely that the Project on Intercomparison of Land-surface Parameterization Schemes (PILPS) will shed light on this issue and provide insight into the accuracy required for adequate representation of land-atmosphere exchanges in GCMs, because this project will in due course evaluate their comparative performance against observational data for selected sites (Henderson-Sellers and Dickinson, 1993).

Pending categorical evidence from PILPS, simple physical insight suggests that many of the SVATS currently available have little substantial difference in the way they

represent the component processes. Assuming this is the case, it is the author's opinion that the creative energy and skill currently being wasted in creating additional SVATS of increasing complexity would be better directed towards providing credible calibration of the few (common) parameters which do matter and, in particular, in seeking ways to rewrite SVATS in (probably simpler) ways, such that these important parameters have enhanced relevance to globally available satellite measurements.

3.2 *The Area-Average Values of Parameters*

If, as suggested above, there are only a few key processes (and associated parameters) which matter in the description of energy and water exchanges between continental surfaces and the atmosphere, the question arises as to how to define the effective, area-average values of these parameters. This is clearly a significant issue for continental, as opposed to ocean surfaces, because the heterogeneity of the land surface is very obvious and occurs at all scales, from local to global. Given that we can separate out the problem of providing parameterization of "mesoscale" circulations in the lower atmosphere induced by surface heterogeneity (Zeng and Pielke, 1995), the primary requirement of a SVAT is to give a simple description of the aggregate, area-average exchanges of a heterogeneous mix of different land covers for each grid square, which is adequate for calculating their influence on the atmosphere in GCMs.

This issue is addressed here head-on and in a way which is consistent with the message of this paper, by advocating that the first model to try is the simplest. That is, adopt (or perhaps simplify) one of the currently available SVATS--providing it includes the key processes and parameters described above--and then define effective area-average values of model parameters to be applied to the vegetation-covered portions of each grid area using a set of simple *aggregation rules* for each parameter (Shuttleworth, 1991, 1992). The philosophy is that, although such simple aggregation rules may not stand the test of time, once in place any critical evaluation of them based on observation or modeling can only result in their replacement by superior alternatives, and progress will automatically follow.

SVATS are required to 'describe area-average surface fluxes, primarily of energy and water, although some attempt is now being made to include description of the exchange of CO₂'. Hence, the required area-average parameters are those which most nearly compute these area-average fluxes, either their integrated value over a given period in the case of stores, or the rate of flow into and out of the stores. In general, the simplest aggregation rule, and the one which is advocated here pending categorical evidence to the contrary, is that *the effective area-average value of land surface parameters is estimated as a weighted average over the component cover types in each grid through that function involving the parameter which most succinctly expresses its relationship with the associated surface flux.*

Let us take the example of momentum flux, τ , and the controlling parameter, z_0 , the aerodynamic roughness. In this case, the relevant relationship between momentum flux and aerodynamic resistance (neglecting other surface parameters) is:

$$\tau \propto \ln^{-2}(z_0/z_d) \quad (1)$$

where z_0 is a hypothetical 'blending height' at some level in the air above the ground where turbulent mixing is sufficient for it to be assumed that the atmosphere has, in effect, lost its knowledge of the differing types of land cover on the ground below. Mason (1988) estimated z_0 to be of the order $L/100$, where L is the characteristic horizontal scale of the vegetation patches with different surface roughness covering the surface. In this case, for a grid square with i component covers of fractional area w_i (perhaps determined from satellite), the relevant aggregation rule for z_0 follows from Equation (1) as:

$$\ln^{-2} [z_0/\bar{z}_0] = \sum_i w_i \ln^{-2} [z_0/(z_d)_i] \quad (2)$$

Table 1 interprets the general aggregation rule given above more generally for this and other key parameters in the land-atmosphere interaction. Certain new symbols are listed in Table 1 in addition to those already defined above. These are specified as follows: C_c is the amount of thermal energy stored in unit area of biomass for each degree change in temperature; S_c is the amount of precipitation which can be stored on the surface of vegetation (and soil) during rain; K_s is the thermal conductivity of the soil; C_s

Table 1. 'Strawman' Aggregation Rules for Key Parameters in Soil-Vegetation-Atmosphere Transfer Schemes (taken from Shuttleworth, 1991; 1993)

PARAMETER	'STRAWMAN' AGGREGATION RULE
Albedo	$\bar{\alpha} = \sum_i w_i \alpha_i$
Maximum Infiltration Rate [Note (i)]	$\bar{I} = \sum_i w_i I_i$
Emissivity	$\bar{\varepsilon} = \sum_i w_i \varepsilon_i$
Aerodynamic Roughness [Note (ii)]	$\ln^{-1} \left[z_s / \overline{z_s} \right] = \sum_i w_i \ln^{-1} \left[z_s / \left(\overline{z_s} \right)_i \right]$
Surface Conductance [Note (iii)]	$\bar{g}_s = \sum_i w_i (g_s)_i$
or [Note (iv)]	$\left(\overline{g_s^{-1}} \right) = \sum_i w_i (g_s)_i^{-1}$
or [Note (iv)]	Average of above two alternatives
Canopy Stores:	
Heat [Note (vi)]	$\bar{C}_c = \sum_i w_i (C_c)_i$
Water [Note (vii)]	$\bar{S}_c = \sum_i w_i (S_c)_i$
Soil Stores:	
Heat [Note (viii)]	$\overline{(K_i C_i)^{1/2}} = \sum_i w_i (K_i C_i)_i^{1/2}$
Water [Note (ix)]	$\bar{S}_s = \sum_i w_i (S_s)_i$

Notes:

- (i) Neglects possible effect of topography on infiltration
- (ii) Averages momentum flux, z_0 is an assumed 'blending height'
- (iii) Assumes very efficient aerodynamic mixing in the atmospheric boundary layer
- (iv) Assumes very inefficient aerodynamic resistance in the atmospheric boundary layer
- (v) Some evidence of adequacy of this assumption (eg. Blyth et al., 1993)
- (vi) Neglects any subgrid variation in radiation (cloud)
- (vii) Neglects any subgrid variation in precipitation
- (viii) Assumes cyclic conduction of soil heat flux into the soil
- (ix) Neglects any impact of topography on water stored in rooting depth (see Section 2.3)

is the specific heat of the soil; and S_s is the amount of water which can be stored per unit area within the vegetation's rooting zone.

Although the 'strawman' aggregation rules given in Table 1 are strictly plausible only in the case of overhead conditions which are constant over a grid square, there is some evidence from modeling studies that they may be more generally valid. It seems wise to consider land surface heterogeneity as two general types according to the length scale of the organization in cover type they contain (Shuttleworth, 1989), because this can influence the atmosphere's response. For Class 'A' land cover, i.e., with vegetation in large patches of about 10 km or greater, conventional mesoscale meteorological models (with intergrid distances of 5-15 km) are a convenient mechanism to test hypothetical aggregation rules. For Class 'B' land cover, which has smaller average patch size (and no obvious surface-linked response in the atmosphere), coupled surface-boundary layer models with intergrid distances of 10-1000 m are appropriate.

Model-based tests of simple, hypothetical aggregation rules, such as those in Table 1, have been made using models of the above two types. Noihlan and Lacarrere (1994) used a mesoscale meteorological model with a 10-km grid to calculate the area-average surface flux exchanges for the Class 'A' mixed cover of the 100 x 100 km, HAPEX-MOBILHY (Andre et al., 1988) experimental square. They regarded these calculated fluxes as the true area-average values, against which they compared estimates of the area-average fluxes made assuming a single one-dimensional SVAT with parameters estimated from simple aggregation rules similar to those defined in Table 1. The results suggest that such rules may be adequate, at least in fine weather. Figure 3 shows a comparison of the two estimates of latent heat which they calculated for one day.

Some similar preliminary tests of aggregation rules have been made for Class 'B' type land cover (actually the most common type) using an atmospheric boundary layer model with short intergrid distances of 100 m. Blyth and Dolman (1993) investigated aggregation rules for aerodynamic resistance and surface conductance using the Penman-Monteith equation as a simple SVAT, with alternate small patches of vegetation with different values assigned to these two parameters. Again, simple aggregation rules work

well in dry conditions, but they found additional complications during and immediately after convective rain storms, when some of the vegetation may be covered by surface water. They found, for instance, that the average of the linear mean and reciprocal mean values for g_a is a more robust aggregation rule in conditions of mixed surface wetness, but that the aerodynamic transfer resistance for sensible heat becomes undefined in such conditions because of the plot-to-plot advection (Blyth and Dolman, 1993).

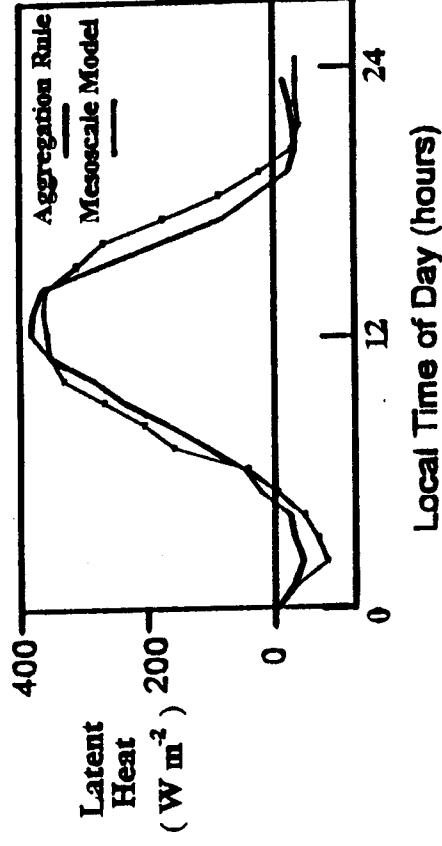


Figure 3. Comparison between area-average evaporation rates calculated for the 100 x 100 km HAPEX-MOBILHY study site for 16 June 1986 using a mesoscale meteorological model to synthesize the hourly values as the sum of the surface fluxes separately calculated for points on a 10-km grid and the value calculated by assuming a single one-dimensional model in which the effective model parameters have been calculated using simple aggregation rules similar to those given in Table 1.

Similar modeling studies have been made by Sellers et al. (1992) using data from the First International Satellite Land Surface Climatology Project Field Experiment (FIFE). In this case, however, the atmospheric part of the (otherwise coupled) model was, in effect, substituted by measurements; the area-average surface conductance was calculated using remote sensing measurement to allow for spatial variability in the light dependence of this parameter. The dependency of g_a on other variables was assumed to be the same or, in the case of soil moisture, measured. This study gives similar good

agreement between calculated area-average fluxes in fine weather but, in this case, the result is given additional credence by the availability of independent aircraft measurements.

Although the modeling studies described above have provided pointers, there is a clear need for more methodical modeling studies of this type, preferably with better focus on the relevant aggregation rules for the more advanced and realistic SVATS. Recent studies of the aggregation rule approach applied within the framework of BATS (Dickinson et al., 1986), using plausible mixes of vegetation covers in a coupled model initiated at selected times with field data from the FIFE study (Arain, 1994), confirm their usefulness in all but artificial situations involving continuously irrigated patches of vegetation in an otherwise dry landscape.

3.3 *Space-Data-Relevance*

There is increasing evidence that the some of the important vegetation-determined parameters used in SVATS may have reasonably robust relationships to variables which can be measured from space. Further, it might be that these relationships are more strongly determined by plant metabolism (whether the plants have a C_3 or C_4 metabolism) rather than the vegetation class to which they are conventionally assigned in advanced SVATS.

Results from the First International Satellite Land Surface Climatology Project Field Experiment (FIFE), for instance, provide encouragement that there may be relationships between one of the parameters which determine canopy resistance and the so-called "Simple Ratio" vegetation index which can be measured from satellite. Following Sellers et al. (1992), the surface conductance, g_s , can be written in the form:

$$g_s = V_p \cdot F_o \cdot [f(\delta e) \cdot f(T) \cdot f(\phi)] \quad (3)$$

where $f(\)$ indicates a 'stress function' for vapor pressure deficit, δe , temperature, T , and soil moisture deficit, j , respectively; F_o is the incident flux of Photosynthetically Active

Radiation (PAR), while σ_F is a factor relating g_s to F_o . Analysis of the FIFE data, which involves first removing the effect of the stress functions to determine ∇_F , reveals that the value of this last factor bears an approximately linear relationship to the Simple Ratio, SR, with the general form:

$$\nabla_F = (SR - a_1) \cdot (a_4 \cdot V_3 + a_2) + a_2 \quad (4)$$

where V_3 is the fractional cover of vegetation with C_3 (as opposed to C_4) metabolism, and a_1 , a_2 , a_3 , and a_4 ($a_4 = 5.668 \times 10^{-5}$) are empirical constants which were determined for the prairie grass at the FIFE site.

It is implied that the constants a_1 , a_2 , a_3 and especially a_4 are strongly determined by the (C_3 or C_4) metabolism of the plants and that the form of Equation (4) and perhaps the value of some of the constants can be more widely applied. It would, however, be naive not to acknowledge that these constants may also in part be determined by the optical characteristics of the leaves. However, because there are fewer metabolic pathways than readily distinguishable biomes, this FIFE result may provide a pointer to the future and suggest that the number of vegetation classes acknowledged in SVATS might reduce over the next few years. Some morphology-related distinctions surely will still need to be made between plants even if they have the same metabolism; perhaps this reduction might just be from the current 15-20 classes to (say) 6-12 classes of land cover.

4. Interfacing Atmospheric and Hydrologic Fields

In Section 2, the case was made for separating the objectives of hydrological modeling in global environmental research as a means of retaining simplicity in the required models. One of the main arguments against making such a separation is that it may inhibit the direct, credible hydrological interpretation of predicted changes in global climate. In this section, it is argued that a more profitable approach is to seek an interface which allows the retention of separation between the predictive and interpretative objectives.

4.1 Nature of the Problem

There can be little doubt that it is through the use of hydrological models that observed variations or predicted changes in global meteorological fields will find most relevant local interpretation in terms of hydrological impact, but the nature of hydrological models is such that they are inherently best framed around geomorphologically determined, irregular river basins. Atmospheric General Circulation Models (GCMs) are primarily required to describe the (largely unrestricted) flow of a fluid medium and, as such, most naturally operate with a rectangular grid which is difficult to relate to underlying irregular watersheds. At first sight, therefore, the information provided by atmospheric models is inherently inconsistent with that required as input to hydrological models.

For computational reasons and at the present time, GCMs operate with a grid mesh vastly greater than that consistent with describing the processes which determine surface and subsurface flows. This complicates use of the meteorological information they provide in the context of worthwhile predictions of the hydrological flows for real catchments. Of more fundamental and lasting significance, however, is the fact that GCMs only require that hydrologically relevant exchanges are described with a level of realism sufficient to allow them to maintain adequate description of globally distributed atmospheric fields of energy, mass, and momentum, because this is their purpose. Therefore, it is cost-effective to formulate hydrologically relevant processes in GCMs only with a level of computational complexity which is broadly consistent with that used elsewhere in the model.

Thus, although the fact that GCMs operate at a vastly different scale to hydrological models is a problem in seeking to provide realistic hydrological interpretation of the meteorological fields, this is primarily a technical limitation which will no doubt reduce in importance with time. The more fundamental and probably more permanent inconsistency between these two types of models arises because of the difference in the relative importance they *a priori* must assign to the continental precipitation and evaporation fields and the inherently different natural geographical elements on which

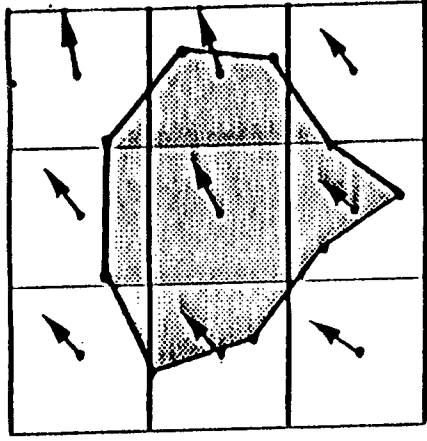
the two types of models operate with most efficiency and realism--rectangles in one case and watersheds in the other.

Some means is required to allow the ongoing development and enhancement of GCMs to proceed in an internally balanced way, but with the need to give them hydrological interpretation still catered for. In this way, GCMs are only required to further one of their own primary objectives, namely that of providing the best possible calculation of the vector field of atmospheric water vapor. Meanwhile, the attention for hydrological modelers need not be on 'cluttering' GCMs with hydrological detail purely for hydrological purposes; rather, it might be on deriving an interface to exploit GCM improvement. The required interface should respond to improving source data, but allow a time series of GCM-generated, global fields of atmospheric water vapor to be interpreted in terms of regional maps of soil moisture and point maps of stream flow discharge at selected locations.

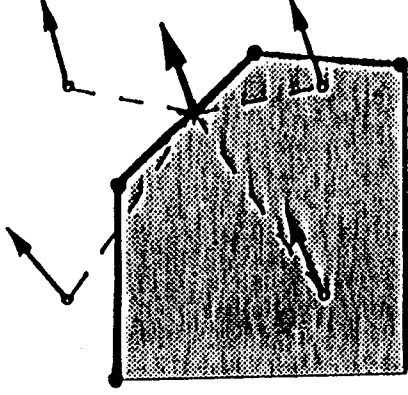
4.2 Reynolds Transport Theorem (RTT)-Based Linkage

For small watersheds with dimensions on the order of 10 km, within-catchment recycling of water is a small component in the water budget and, at this scale, the input required for hydrological modeling is the net input of water across the small catchment. Given such an input for each individual small basin in a larger watershed, conventional and now well-established routing procedures can be applied. In many cases, these can exploit an existing locally relevant calibration. Surface observations are the normal source data for calculating the area-average net water input to small catchments, but the vector fields of atmospheric water vapor generated by GCMs are an alternative source of net water input, at least for the purpose of calculating monthly or annual flows. The Reynolds Transport Theorem can be applied to that portion of the atmosphere which overlies each catchment to provide the required interface.

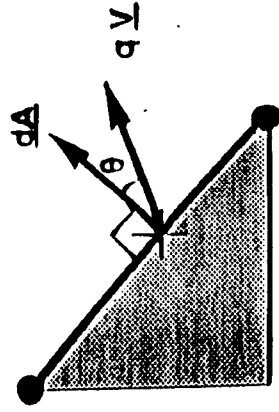
Discharge is only defined for irregularly shaped watersheds (Figure 4a). It is possible to define a vector field of atmospheric moisture on the GCM grid by taking the product



(a) A watershed overlaid on a GCM grid

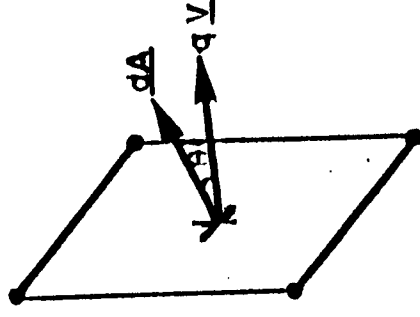


(b) Interpolation of the moisture flux vector onto the watershed boundary

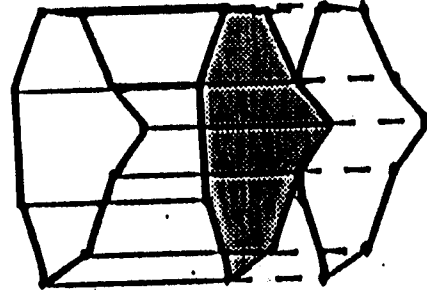


$$dQ_a = q \cdot V \cdot dA \\ = q \cdot V \cdot dA \cdot \cos \theta$$

(c) Computation of the moisture flow through the boundary



(d) The boundary as a vertical plane in the atmosphere



Atmospheric water

Surface water

Subsurface water

(e) The watershed as a 3-D control volume

Figure 4. Illustrating steps in the development of a Reynolds Transport Theorem based interface between the vector field of atmospheric humidity provided on a regular rectangular grid by a GCM (or NWP model), and the net input of water to an underlying array of irregularly shaped watersheds for which traditional hydrological modeling is required.

of the specific humidity, q , and the velocity components u and v for each position in the grid box, integrating the values qu and qv vertically through the atmosphere, and then averaging them through time if necessary. The values of qu and qv together form the moisture flux vector, $q\mathbf{V}$, and can be defined anywhere within the domain of the vector field by spatial interpolation between the four most adjacent point values. In particular, they can be defined at the center of each line segment in the watershed boundary, as in Figure 4b. (Note: To simplify the explanation here, the linear dimension of the rectangular grid in Figure 4a has the same order of magnitude as that of the schematic watershed. With present day GCMs, this is not the case, but this fact does not alter the proposed procedure in any fundamental way. It is, however, likely to degrade the quality of the product provided by an RTT-based interface pending a reduction in the GCM grid size on which meteorological fields are calculated. This could give rise to the need for grid-size dependent corrections in the short term, as discussed in Section 4.3).

If the line segment of the watershed boundary is projected vertically from the ground to the maximum height used in the original GCM field, it defines a surface area which we represent as a vector $d\mathbf{A}$, which is defined to be normal to the area always pointing in the outward direction from the watershed, as in Figure 4c and Figure 4d. The contribution to atmospheric moisture flow above the catchment through this surface, dQ_s , is given by:

$$dQ_s = q\mathbf{V} \cdot d\mathbf{A}$$

or:

$$dQ_s = qVdA \cos\theta \quad (5)$$

where θ is the angle between the vectors $\mathbf{V}$ and $d\mathbf{A}$. Because the length and orientation of all line segments are known, θ can be determined and dQ_s calculated.

In this way, the watershed can be visualized as a three-dimensional control volume by similarly projecting vertically all the line segments along its boundary and then enclosing the volume with horizontal planes at the top and bottom, as shown in Figure 4e. The net

atmospheric moisture flow out of the watershed is then found by integrating Equation (5) over all the planes making up the control surface, thus:

$$Q_a = \oint_{c.s.} qV \cdot dA \quad (6)$$

Once this method has been successfully implemented for a single watershed, its application could be automated using a Geographical Information System, and it could then be applied to a landscape which may comprise hundreds or even thousands of watersheds if necessary. In this way, it is possible to conceive the watershed as a fundamental spatial primitive and to use it to describe the hydrological systems in a landscape in much the same way that the rectangular grid cell is the fundamental spatial primitive is used to describe atmospheric flow over the Earth's surface.

The net atmospheric moisture flux upwards from the watershed surface is easily found as:

$$q_a = Q_a / A_w$$

where A_w is the watershed area, so that:

$$q_a = \frac{1}{A_w} \oint_{c.s.} qV \cdot dA \quad (7)$$

and $q_a = E - P$ where E is evaporation and P is precipitation. The water balance on the watershed is hence given by:

$$\frac{dS}{dt} = -Q_a - Q_s - Q_g \quad (8)$$

$$\frac{dS}{dt} = -(E - P) A_w - Q_s - Q_g$$

where Q_s , Q_p , and Q_g represent the outflows of atmospheric water, surface water, and groundwater across the control surface, respectively, and S is the total of all water storage components in the watershed.

The above logic, the Gauss theorem, can be used to relate flux divergence within a control volume to the integral of flux over the control surface. However, if we wish to examine the time rate of change of a fluid passing through a control volume, it is necessary to apply the Reynolds Transport Theorem (Aris, 1962), which incorporates the Gauss theorem along with basic physical laws and applies them to the fluid-flowing through a control volume at a fixed location. This can be done as follows.

If B is the amount of a fluid property, b is the amount of this property per unit mass of fluid, ρ is the fluid density, and dv is an element of volume within the control volume, then the Reynolds Transport Theorem, when applied to a control volume of fixed size such as that shown in Figure 4e, takes the form:

$$\frac{dB}{dt} = \frac{d \left[\iiint_{c.v.} (\beta \rho) dv \right]}{dt} + \oint (\beta \rho) \mathbf{V} \cdot d\mathbf{A} \quad (9)$$

where the term dB/dt is the material derivative of the fluid property (that is, its total rate of change with respect to time); while the first term on the righthand side is the time rate of change of the amount of property stored in the volume, and the second term is the net outflow of the property across the control surface. When this theorem is applied to water vapor transport in air, B becomes the mass of water vapor, β the mass of water vapor per unit mass of moist air, (i.e., the specific humidity q), ρ the density of moist air, and dB/dt becomes the sum of the rate of extraction of vapor from the air stream by precipitation and the rate of addition to the air stream by evaporation, respectively.

If the flow is assumed to be in steady state, an assumption which can be argued to be valid for monthly average calculations, the first term on the righthand side of Equation (9) disappears and the physics in the equation reduces to that used to derive Equation (6). Alternatively, when studying short-term atmospheric moisture motion, such as might occur during thunderstorm development over a small watershed, the moisture content of the air changes rapidly. It is then necessary to take account of all three terms in the Reynolds Transport Theorem. In this case, the first term on the righthand side of

Equation (9) can be accounted for by interpolating the vertically integrated values of specific humidity into a subgrid across the watershed, areally integrating these values at each point in time, and then approximating the time derivative of atmospheric water vapor from the successive differences between these integrated quantities. Exactly the same procedure can, of course, be used in the case of monthly average calculations to account for longer term changes in water vapor concentration.

4.3 Short-Term Weakness and Upcoming Opportunity

GCMs are used in two main ways. They can be run as free-standing, self-regulating, predictive models, and it is in this form that they are used as the basis of global change prediction. However, in the form of Numerical Weather Prediction (NWP) models, they are routinely run at regularly repeated intervals, typically every few hours, to simulate the evolution of global fields for hours and days ahead, and they are then the basis of weather prediction. Prior to each short run, meteorological data from around the world in the form of surface and balloon borne measurements and, if feasible, satellite measurements are *assimilated* using advanced error balancing techniques into the last predicted global field calculated by the model. This field is thus modified in a model-consistent way so that the initial state from which the next prediction is then made is as close as possible to reality, within the limits of the data available. The, albeit in part model-determined, global fields created by assimilating all the available meteorological data to hand at the time of re-calibration are in themselves a source of meteorological data. These data are by their nature available worldwide and, being model-generated, have the advantage of being consistent with the constraints of energy and mass imposed in the model.

The time sequence of evolving global meteorological fields generated by the means just outlined is sometimes referred to as *four-dimensional (4-D) model assimilation data* and is routinely stored by the major numerical weather prediction centers several times each day as a byproduct of their operation. As stated above, although substantially determined by the assimilated input data, they are also significantly influenced by the

form and assumptions employed in the model used in their synthesis. They need to be used with understanding and care because of this.

A point of particular relevance to the interface outlined above is that the information on net water exchange with the ground derived by the Reynolds Transport Theorem calculations can only be no more than that contained in the original meteorological source data. These data provide average values of this surface exchange sampled at the spatial interval of the grid used in the GCM (or NWP) model, and the surface fluxes calculated by the interfacing method we derive can, therefore, be no more than smoothed grid-average values. Thus, it is probable that there will be differences between the RTT-calculated net input of water and the measured runoff for individual catchments whose size is less than that of the grid used in the atmospheric source data.

The fact that an RTT-based interface can at present provide no more than smoothed (GCM or NWP) grid-average values of surface exchange over small watersheds will be of most significance in regions with marked topography. With atmospheric humidity fields derived from large grid mesh GCMs, it is likely that there will be substantial differences between the RTT-calculated net input of water and measured runoff for individual hilly catchments. Further, although these discrepancies will likely decrease with catchment size, it is possible that some systematic discrepancy will persist.

On the basis of the above comments, it is of interest to determine whether discrepancies in RTT-based calculations with present-day GCM fields originate from:

- (a) inherent regional weaknesses in the GCM-calculated source data with regard to its description of large-scale humidity fields;
- (b) scale-dependent inconsistencies between the meteorological source data and the catchment scale to which it is applied; or
- (c) shortcomings in the GCMs physics associated with poor recognition of topography-generated vapor extraction arising from processes which occur at subgrid scales.

However, there is opportunity to address evaluation of these potential discrepancies in the course of the next few years.

Activity under the GEWEX Continental-scale International Project (GCIP) will provide upgraded global fields of meteorological data as part of the enhanced observational program which will run for five years from 1995. Numerical Weather Prediction models at the National Meteorological Center (NMC) in Washington and at the European Centre for Medium Range Weather Forecasting (ECMWF) in Reading, UK will be among the operational models used primarily for this purpose. These centers already have model-calculated data records for an extended period which were generated in the course of their day-to-day operational activities and which are available for research purposes. Both centers have plans to upgrade the continuity of these global fields by re-analyzing, with a consistent model applied to the assimilation of the original source data used in past assimilations.

In parallel with this improvement in meteorological data under GCIP, there will be an equivalent synthesis of hydrological data (WMO, 1992). Specifically, it is proposed to define a substantial set of watersheds sampled from across the Mississippi River basin and to prepare a high-quality data set of rainfall and runoff for these in the near future. These data could be used for initial comparison against calculated net water inputs made using the Reynolds Transport Theorem interface for these basins. The extensive nature, range of climates, and number of available catchments will ensure the provision of a worthwhile initial test of the interfacing methods and will provide an ample source of information for investigating the relationships between catchment characteristics and any differences in calculated net water input to the catchment and measured runoff.

Presumably, the area average values at larger scales ought to agree better and, in the context of GCIP, an investigation could be made to establish the scaling characteristics of the RTT-based interface. This would involve applying it to progressively larger catchments in the Mississippi River basin, each being comprised of smaller watersheds to which the method can also be applied and for which runoff data are available. It is to be anticipated that there will be substantial 'noise' in the relationship between RTT

calculations and observed runoff at the small basin scale even in monthly average values, but that this will progressively reduce and then stabilize as the dimensions of the sampled watershed approach and then exceed those of the GCM grid.

Finally, it should be remembered that the Mississippi River basin is in fact a comparatively flat catchment. Therefore, additional evaluation of the performance of the RTT-based interface is required outside the strict geographical limits of the GCIP study area to investigate potential topography-generated discrepancies. Watersheds with marked topography in the western USA are clearly the most obvious target study sites.

5. Concluding Comments

The central objective of this paper is to provide guidance and perspective, but also reassurance and motivation to hydrometeorological modelers who are faced with the task of applying their skills, experience, and process understanding to the task of improving and applying global environmental change predictions.

My central message is that, in approaching this challenge, we should use our physical insight to distinguish between the objectives we need to address and define what model components truly matter in order to fulfill those objectives separately and with minimum complexity. We cannot, and need not, apply small-scale hydrological modeling skills mechanically and universally across the globe, nor do such models need to be absorbed into GCMs in their present complex form and run in real time. If we recognize this, our task becomes tractable and, returning to the theme of this workshop, we can approach 'the trials and tribulations of hydrological models and measurements' in the context of GEC with increased vigor and focus.

6. Acknowledgments

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Surface Flux Measurement and Modeling at a Semi-Arid Sonoran Desert Site

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Abstract

Continuous measurements of standard meteorological variables using an automatic weather station and intermittent measurements of the surface energy balance, carbon dioxide flux, and momentum flux using Bowen ratio, eddy covariance, and sigma-T instrumentation were made for 13 months at a semi-arid Sonoran Desert site just west of Tucson, Arizona. Weather observations demonstrate typical semi-arid Sonoran desert conditions, with frequent clear skies, high radiation, a large seasonal and diurnal temperature range, low relative humidity, and intermittent precipitation mainly of convective origin during a summer monsoon season. The substantial observational problems associated with surface flux measurements in this environment are reported. Comparisons between measured fluxes made simultaneously with different instrumental systems show acceptable agreement. Most of the incoming radiant energy leaves as sensible heat, and latent heat fluxes are always low, but transpiration is enhanced for about 10 days after rain. To investigate the influence of Crassulacean Acid Metabolism plants on carbon dioxide flux, measurements were sustained through the night. Carbon dioxide uptake is low, typically with peak daytime uptake in the order $0.25\text{--}1.0\ \mu\text{mol m}^{-2}\text{ s}^{-1}$ for the period for which data are available, and some carbon uptake persists even at night. The observations were used to validate and calibrate the surface energy balance simulated by the *Biosphere-Atmosphere Transfer Scheme*. Using the default "semi-desert" soil and vegetation parameters specified in the National Center for Atmospheric Research *Community Climate Model version 2* resulted in a poor simulation of observations. However, using a set of site-specific parameters, including on-site observations to specify more realistic soil and vegetation characteristics, and optimized minimum surface resistance and plant wilting parameters, resulted in a substantial improvement in model performance. The site-specific parameters reflect the fact that the vegetation fraction is greater than assumed in the default parameter set, that leaf area index and minimum stomatal resistance are less, soils at the study site contain more clay, but that the plants' wilting point is lower than this clay fraction would imply. The modified, site-specific parameters more accurately describe the conservative character of the semi-desert vegetation and the moderate nature of its response to the seasonal water cycle.

1. INTRODUCTION

Semi-arid environments cover 40% of the Earth's land surface and therefore are a significant component of the Earth's climate system (Moran *et al.*, 1994). The dominant forcing in semi-arid regions is precipitation, which can be extremely variable in both time and space, resulting in great variability in soil moisture and in latent and sensible heat fluxes (Stewart *et al.*, 1994). The extreme seasonality of precipitation in semi-arid regions, such as the July-August Sonoran desert monsoon has a profound effect on seasonal patterns of surface water and energy exchange. In fact, the seasonal patterns in surface fluxes caused by monsoon precipitation greatly exceed those caused by the seasonal differences in solar forcing.

During long dry periods, the surface-atmospheric exchanges of water and energy are controlled by a complex combination of soil water availability and perhaps unique plant physiology. During short wet periods, evapotranspiration approaches its potential, and is not limited by soil water availability, but during dry periods, low soil water availability limits surface evaporation severely. Soil water availability also limits plant transpiration, but several different metabolic pathways exist in semi-arid plants that allow them to sustain photosynthesis when soil water is very low. Succulent plants use the Crassulacean Acid Metabolism (CAM) pathway that conserves water by only transpiring at night and utilize in-plant water storage to become independent of soil water availability, others develop extremely deep roots that tap groundwater reservoirs, while others shed leaves and continue photosynthesis in their stems. Understanding the complex interactions between plant physiology, extreme seasonality in precipitation and solar forcings, and large diurnal temperature variations in semi-arid environments is crucial for assessing their role in the Earth's climate system. Further, semi-arid regions are climatologically sensitive because slight shifts in global climate could potentially result in major changes in local surface conditions. Despite this, surprisingly little is known about the spatial and temporal variation in energy and water fluxes and how these are influenced by vegetation and soil properties in semi-arid environments.

In order to understand the complexities of the semi-arid soil-vegetation-atmosphere system which determines exchanges of water, energy, and carbon dioxide and eventually predict the interrelationship between this fragile land cover and climate, several field experiments, including the Sahelian Energy Balance Experiment (SEBEX; Wallace *et al.*, 1991), MONSOON-90 (Kustas *et al.*, 1991), the Owens Valley study (Kustas *et al.*, 1989), the Smith Creek Valley, the Smoke Creek Desert studies (Nichols, 1992), and the La Crau observational study (Kohsiek, *et al.*, 1993) were undertaken. These experiments provided valuable insight, but have generally been of short duration and have focused mainly on investigating spatial, as opposed to temporal, variability. This study focuses on measuring and modeling the surface fluxes from a pristine semi-arid environment over a year, in order to better characterize the climate-related signals in the surface fluxes.

Soil-Vegetation-Atmosphere Transfer Schemes (SVATS) are one-dimensional submodels used in Global Circulation Models (GCMs) to describe the interaction between the overlying atmosphere and the vegetation and soil. The validity of the description provided by SVATS has not been fully explored in semi-arid regions. In response to this need, the performance of one of the

most commonly used SVATS, the Biosphere-Atmosphere Transfer Scheme (BATS; Dickinson *et al.*, 1993) is evaluated using the surface energy and water fluxes collected in this study.

2. MATERIALS AND METHODS

2.1 Experimental Site

Field data were gathered at a site located at 32°13' N, 111°5' W in the semi-arid, alluvial Sonoran Desert near Tucson, Arizona, on gently sloping terrain at an elevation of 730 m. There are uninterrupted fetches of many kilometers in all directions except 1 km to the northeast, where the 1340 m high Tucson Mountain range lies. Total precipitation measured over the year-long sampling period starting on May 12, 1993 was 275 mm.

Vegetation was interspersed with patches of exposed rocky soil, giving a fractional vegetation cover of 40%, estimated from four linear 30-m transects within the fetch of the Bowen ratio instrument (Woodhouse and Zeng, 1993). Key species include creosote bush (*Larrea tridentata*), brittle bush (*Encelia farinosa*), triangular leaf bursage (*Ambrosia deltoidea*), velvet mesquite (*Prosopis velutina*), palo verde (*Cercidium microphyllum*), ironwood (*Olneya tesota*), jojoba (*Simmondsia chinensis*), ocotillo (*Fouquieria splendens*), non-native agave (*Agave americana*), as well as various cacti including saguaro (*Carnegiea giganteus*), Saguaro, chollafruit and teddybear cholla (*Opuntia versicolor*, *O. fulgida* and *O. bigelovii*, respectively), and Englemann's prickly pear (*Opuntia phaeacantha*). Of these, all the succulent species (cacti and agaves) employ the Crassulacean Acid Metabolism (CAM) while the other species employ the more common C3 pathway for photosynthesis. For short periods in the spring and after the summer monsoons, a sparse cover of short black gramma grass (*Bouteloua eriopoda*; C4) and flowering annuals developed on the otherwise exposed soil. Vegetation heights range from a few tens of centimeters for low grasses and bushes up to 7 m for the tallest saguaro cacti; mean vegetation height was estimated as 1.2 m.

Soil samples were obtained at depths of 1 cm and 15 cm from a bare soil area, a vegetated area (under a bush), and a stream bed. These samples were sifted through a D₁₀ mesh sieve, and the fine fraction was analyzed for soil texture and albedo. The coarse fraction, consisting mainly of rocks, had a similar color to the fine fraction taken from the same sample. In fact, samples taken from the bare and vegetated surfaces are more representative of the average soil parameters for the field site than those from the stream bed, because the latter constitute just a small fraction of the surface area. All sites had a significant fraction of clay, with the least clay fraction in the stream bed samples. The soil texture was assigned to one of the BATS soil classes (see Dickinson *et al.*, 1993, Table 3) on the basis of porosity; overall, the BATS soil texture class 9 was the most representative of the soil at the field site. For the near-surface samples, soil albedo measurements for wavelengths $\lambda < 0.7 \mu\text{m}$ indicated albedo ranges of 0.19 to 0.21 and 0.093 to 0.11 for dry and wet samples, respectively, corresponding to BATS soil color index 3 (see Dickinson *et al.*, 1993, Table 3).

After several weeks without precipitation, on September 23, 1993, a bulk soil density of 1700

kg m⁻³ was determined from soil samples taken at the field site, when the gravimetric soil water content was 0.08 kg water per kg soil (volumetric water content, 0.136 m³ of water per m³ of soil).

2.2 Micrometeorological Instrumentation and Methods

The instrumentation used in this study is shown in Figure 1 and includes a Bowen ratio system, sigma-T systems, an eddy covariance system, and a standard meteorological station. The Bowen ratio system measures the height-dependent difference in temperature and humidity at two levels and derives the sensible and latent heat fluxes via an energy balance calculation. The sigma-T system measures the standard deviation of air temperature using fast response thermocouples to provide an alternative estimate of sensible heat flux. The eddy covariance system uses an infrared gas analyzer and a sonic anemometer to provide direct measurements of sensible and latent heat, momentum, and carbon dioxide fluxes. Finally, the automatic weather station provides routine measurements of precipitation, net radiation, incoming short-wave radiation, soil heat flux, air temperature, relative humidity, and wind speed and direction.

2.2.1 Automatic Weather Station

Standard meteorological measurements were taken using an automatic weather station (*Campbell Scientific*, Utah, USA) from May 12, 1993 to June 5, 1994. Measurements of net radiation, incoming short-wave radiation, air temperature, relative humidity, wind speed and direction, precipitation, soil temperatures, and soil heat flux were recorded on a data logger, initially as average values over 10-minute intervals. These were subsequently combined to provide time averages over longer intervals, as required. Recognizing that about half the land surface is vegetated, net radiation was taken as the arithmetic average from two net radiometers (*R.E.B.S.*, Washington, USA; model Q-6), situated to sample the extremes of net radiation variability due to vegetation. One was mounted at a height of 3.4 m above a palo verde bush, and the second at a height of 2.9 m above a mixture of bare soil and small sagebrush plants. The instrumental error associated with these net radiometers is estimated as 5%, while the systematic difference associated with the underlying ground cover was greater than this, the daytime measurement over the short vegetation and soil being typically 10-12% less than that over the bush.

The pyranometer (*Li-Cor*, Nebraska, USA; model LI-200SZ) and a combined thermometer and hygrometer (*Campbell Scientific*, Utah, USA) were respectively mounted 8.3 m and 6.7 m above the ground. Factory calibration specifies a maximum absolute error of 5% for the pyranometer, but suggests typical values of 3% under natural daylight conditions. At 20 °C, the accuracy of the hygrometer quoted by the manufacturer is 2% for the relative humidity (*RH*) range 0-90% or 0.3 g/kg for the specific humidity (*q*) range 0-14.5 g/kg, and 3% for the *RH* range 90-100% or 0.5 g/kg for the *q* range 14.5-16.1 g/kg. Temperature measurement accuracy is 0.4 °C for the temperature range -33 to +48 °C. The anemometer and the wind vane, with estimated measurement accuracies of 2% and 5%, respectively, were installed on top of the 9-m tower. Precipitation was measured using a tipping bucket rain gage (with funnel top 31 cm above the ground), which had a quoted accuracy of 1% at rainfall rates of 50 mm h⁻¹ or less.

Following the same methodology used for averaging the net radiation measurements, the areal average soil heat flux was obtained as an average of fluxes measured at two bare-soil and two vegetation-covered sites, recognizing that about half the land surface is vegetated. Four soil heat flux plates were buried 8 cm below the soil surface, two beneath vegetation and two under bare soil. The average soil temperature above the heat flux plates was determined by averaging soil thermocouple measurements at 2 cm and 6 cm above the plates. The surface heat flux was calculated by adding the measured heat flux to the energy stored in the layer above the heat flux plates, the latter being proportional to the rate of change of soil temperature. The estimates of the thermal capacity of the soil required for this calculation (Woodhouse and Zeng, 1993) were based on fairly dry soil samples obtained on September 23, 1993 (see Section 2.1 above). This estimate of thermal capacity is valid throughout most of the measurement period, except during some short wet periods where the lack of soil moisture information may lead to errors in soil heat flux. Sensitivity tests revealed that likely changes in the thermal capacity for moist soil during and immediately after rainstorms give only a small ($\sim 2\%$) error in the measured surface heat fluxes.

2.2.2 Eddy Covariance System

The eddy covariance system used in this study was assembled following the design of Moncrieff *et al.* (1995) and implemented to operate in an isolated field environment using solar energy. It used a 3-axis ultrasonic anemometer (*Gill Instruments*, Fants, UK; Model 1012R2A) to measure wind speed variations and air temperature, a carbon dioxide and water vapor infrared gas analyzer (*Li-Cor*, Nebraska, USA; model 6262) to measure variations in concentration, and a high-speed laptop computer for system control and data processing, together with an air ducting system (including high-speed pumps and chemical air scrubbers), and other assorted data handling and power-related hardware. The computer made measurements of fluctuating variables 20 times per second.

The sonic anemometer and gas analyzer intake were situated well above the vegetation on top of a 6-meter tower. Because the sonic anemometer was of a 3-axis design, it allowed measurement of the three-dimensional wind vector, and it was deployed in an operational mode in which it makes manufacturer-provided real-time corrections for the flow distortion and wind shadowing generated by the anemometer structure. The anemometer was linked to the portable computer through a serial port and had the option of monitoring additional voltage inputs. These were used to collect the concentration measurements from the infrared gas analyzer.

The gas analyzer measured carbon dioxide and water vapor concentrations at high speed by using the differential absorption of infrared light by these two gases. It uses a light chopper to allow rapid, alternate samples of the light absorption by a zero concentration reference gas to correct for detector drift. The gas analyzer was calibrated at least once each week against gases with known concentrations of water vapor and carbon dioxide, these being derived from a dew-point generator (*Li-Cor*, Nebraska, USA; model LL-610), a compressed reference gas (400 ppm CO₂ accurate to 1% of the National Institute of Standards standard), and "scrubbing" chemicals to provide air samples with zero concentrations. During measurements, air was ducted 10 meters (using "Bev-a-Line IV"

polyethylene tubing) from the vicinity of the sonic anemometer, down the tower and into the infrared gas analyzer, using a high-speed pump operating at a flow rate sufficient to assure turbulent flow in the ducting tubes. A pressure sensor was included in the gas flow path adjacent to the sampling chamber to allow correction for pressure changes associated with air movement through the ducting system.

The eddy covariance system consumed approximately 60 W of power, mostly by the pumps and the infrared gas analyzer. This power was supplied by ten (46 by 91 cm) solar panels regulated to charge ten 105 amp-hour deep-cycle marine batteries. Power supplied to the gas analyzer and pumps was regulated by DC-DC converters, use of which also helped to prevent "ground loop" problems. In this application a DC-AC inverter was used to supply the computer, tape drive, and sonic anemometer with 110 volts AC because the DC-DC converters suffered momentary power interruptions which stopped data collection. Solar power was plentiful at this site, but high daytime air temperatures and heat produced by instrumentation lead to instrument damage, necessitating the development of an instrumentation enclosure with enhanced air circulation, and high reflectance properties.

The process of ducting an air sample from the intake near the sonic anemometer on top of the tower to the gas analyzer introduces a delay of several seconds between the wind vector and concentration measurements. This must be allowed for during data analysis. The on-line computer was not powerful enough to do the required processing in real time, so the raw wind vector and concentration data were saved for retrospective analysis. About 20 megabytes of data were stored per day, then transferred and reprocessed. Ultimately, the volume of data and the complexity of retrospective data processing (see below) associated with the prolonged use of this eddy covariance system provoked a decision to choose the alternative Bowen ratio system for the year-long routine flux measurements undertaken in this study.

Retrospective processing of the stored eddy covariance measurements was a two-step process. The air-ducting delay time was determined using a cross-correlation analysis between the measurements of carbon dioxide and water vapor and air temperature as measured by the sonic anemometer. However, in this environment where measured fluxes are often low (especially at night), the delay time is not always easily identifiable from an individual cross-correlation analysis made at a particular time, and a process of manual interpolation was required to select the delay times (so justifying the storage and retrospective processing of the data). This re-analysis involved plotting the time series of cross correlation spectra for each 20-minute sampling time and assuming that delay time was fairly consistent from one time period to the next. Delay times were relatively uniform, with changes typically of less than about one second in a day but sometimes with changes up to several seconds in a week.

The second step in the analysis was to apply calibration to the raw voltages and to calculate the covariances, variances, and means from the several variables using the time alignment between the data streams associated with flow down the ducting tubes derived as above. The "EDDYFLUX" software (Verhoef, 1992) was used for this purpose. This software performs a coordinate rotation

analysis as well as making corrections for sensor response, path length averaging, sensor separation, dampening of fluctuations and contamination of CO_2 flux by latent heat flux (Moore, 1986; Philip, 1963; Leuning and King, 1991; Webb *et al.*, 1980), for all of which a correction exceeding 5% is rarely needed. Finally, a correction to account for the change in carbon dioxide storage below the sensor was made, where it was assumed that the carbon dioxide concentration below the sensor was equal to the concentration measured at the sensor.

2.2.3 Sigma-T Measurements

Tillman (1972) proposed a simple procedure for estimating sensible heat flux which is based on Monin and Obukhov's (1954) similarity theory applied in unstable atmospheric conditions and which has become known as the "Sigma-T" method. In very unstable conditions well above the ground, it requires measurements only of the standard deviation of temperature, σ_T , and the mean air temperature, θ , at height, z , together with an estimate of the zero plane displacement height, d . The sensible heat flux is estimated from:

$$H = \rho_a c_p k \left[\frac{\sigma_T}{C_1} \right] \frac{kg(z-d)}{\theta + 273.2} \quad (1)$$

where k is the von Karman constant (≈ 0.4), and C_1 is a constant which is often set to 0.95 following Tillman (1972). Because this method is only valid during unstable atmospheric conditions, it cannot be applied between dusk and dawn (*i.e.*, during the night), or during rain, because the atmosphere is usually neutral or stable at these times, or during rain because evaporation from the sensors produces invalid standard deviations.

In this study, the measurements required to apply this technique were made using fine-wire (76 μ m diameter) thermocouples. These were installed on arms extending 0.75 m from the tower, at a height of 7 m from June 9, 1993, and also at 5 and 10 m between November 1 and December 31, 1993. The standard deviation relative to a running mean and the arithmetic average of air temperature were recorded on a data logger over 20-minute intervals, with a resolution equivalent to 0.006 °C at a frequency of 10 Hz.

2.2.4 Bowen Ratio Measurements

The Bowen ratio-energy balance method relies on measuring the components of the surface energy budget

$$H + L_e = R_n - G \quad (2)$$

Here the net radiation, R_n , and the soil heat flux, G , are directly measured; while the ratio of L_e and H are estimated from measurements of the difference in vapor pressure, de , and potential temperature, dT , between two levels above the ground. The Bowen ratio, β , the ratio of the sensible

to the latent heat flux, is assumed to be proportional to these differences, thus:

$$\beta = \left(\frac{c_p p}{e \lambda} \right) \left(\frac{dT}{de} \right) \quad (3)$$

where:

- p = Atmospheric pressure (kPa)
- c_p = Specific heat of air ($\text{kJ kg}^{-1} \text{ } ^\circ\text{C}^{-1}$)
- λ = Latent heat of vaporization (kJ kg^{-1})
- ϵ = Ratio of molecular weight of water to that of air.

Combining the definition of the Bowen ratio with the energy-balance equation gives:

$$L_e = \frac{(R_n - G)}{(1 + \beta)} \quad (4)$$

with the sensible heat flux then derived as the residual in the energy-balance equation as:

$$H = R_n - G - L_e \quad (5)$$

Net radiation and ground heat flux were measured for the period May 12, 1993 to June 5, 1994, with measurements of the Bowen ratio also attempted over this same period using a proprietary hardware system (*Campbell Scientific*, Utah, USA). Temperature and humidity sensing was carried out near the end of two 1.5-m long arms, one mounted near the top of the 10-m high tower and the other 3 m above the ground. Air temperatures were measured using 76 μm diameter chromel-constantan thermocouples, with the differential voltage output between the two levels monitored with a data logger resolution equivalent to 0.006 $^\circ\text{C}$. The vapor pressure was measured with a cooled dew-point hygrometer (*General Eastern Corp.*, Massachusetts, USA; model DEW-10). This hygrometer alternately sampled air ducted from intakes adjacent to these two thermometers every two minutes, via (1 μm pore size) Teflon filters and (Bev-A-Line IV) polyethylene tubing, with a 2-liter buffering container in each air line. Dew-point measurement errors are estimated as 0.5 $^\circ\text{C}$, corresponding to better than 0.01 kPa in vapor pressure resolution over most of the range -10 $^\circ\text{C}$ to 70 $^\circ\text{C}$ ambient temperature. The difference in humidity between the two levels is small in the semi-arid environment, and the original separation of the two arms (3 m) proved inadequate to resolve the (humidity) gradients. Even the greater separation (7 m) used after July 6, 1993 provided only marginally adequate resolution, as described below.

In practice, the Bowen ratio system provides only intermittent measurements because of hardware problems, these being particularly severe at the outset of the study. These problems are outlined here for the guidance of others planning to operate such a proprietary system in a similarly exacting semi-arid environment. Although equipment maintenance was reasonably simple,

substantial data were lost due to small air leaks produced by ultraviolet degradation of the polyethylene ducting tubes which can subtly compromise the measurement. Such tubing exposed to intense sunlight should be changed periodically to prevent leakage. Other hardware problems encountered in this study include the occasional sticking of the solenoid valve used to switch the air flow sampled by the dew point sensor and damaged thermocouple junctions due to bird activity and hailstones.

However, the most troublesome aspects of this particular Bowen ratio measuring system when operating in a semi-arid environment are associated with the cooled dew-point mirror. Under conditions of very low humidity, and with dew point commonly well below 0 °C, the normal operating range of this mirror is easily exceeded. The device then sometimes fails to sense that the proper dew-point had been reached during a cooling cycle, and the heat pump continues cooling the mirror, causing persistent ice formation. This condition, if not detected in time, may cause the mirror to chip or scratch and, if the ice is allowed to build up for more than a few days can (and in our case did) cause heat pump failure. Ready access to a spare dew-point mirror and cooling system is therefore advisable when using this proprietary system in semi-arid environments. A heat source (an automotive light bulb connected to a thermostat set to turn on when the temperature in the box dropped to 10 °C) was installed near the dew-point mirror block in the enclosure box. This alleviated the freezing problem to some extent by speeding recovery from persistent icing.

2.3 Analysis Procedures and Quality Control

The near-surface weather variables obtained from the automatic weather station (AWS) for the first 365 days of the study (from 12 May 1993 to 11 May 1994) were used in the modeling component of this study. There were no missing data in the AWS record; however, occasional measurements from one of the two net radiometers were unreliable because of bird damage to radiometer domes. Normally the preferred value of net radiation is the average value from the two net radiometers, but when one was unavailable, a value for the missing measurement was derived from the other using the linear regression between the two radiometers established at other times. Soil heat flux was also normally calculated as the average of the two sets of soil heat flux instruments, but again, occasionally missing data records were allowed for by exploiting the linear regression derived when both systems were available.

Only a proportion of the original raw Bowen ratio data were considered reliable enough for use in the BATS validation exercise due to various mechanical failures and other system limitations, which were most severe during initial data collection. Of the 389-day period for which data collection was attempted (from May 12, 1993 to June 5, 1994), only 170 days were determined to contain some valid data, these falling within the period August 10, 1993 to March 27, 1994. However, even within this subset of days, not every 20-minute sampling period was considered acceptable; in fact, only 30% of the measured vapor pressure and temperature differences were considered reliable.

A deliberately exacting set of criteria was applied to select among these Bowen ratio data to

ensure their credibility, the primary exclusion being to ignore data when the observations were considered beyond the instrumental accuracy of the Bowen ratio system as a whole or the individual sensors involved in that system. Accordingly, observations in which the absolute value of the vapor pressure difference between the two measurement levels, $|d_{el}|$, was less than 0.005 kPa were excluded, as were observations for which the Bowen ratio was close to -1, specifically for the range $|1 + \beta| < 0.3$. The latter occurs when sensible and latent heat fluxes are in opposite directions and approximately equal, when the Bowen ratio method cannot determine the size of the surface fluxes. This condition routinely occurs for short periods around dawn or dusk when the energy available for evaporation (and both surface energy fluxes) is low and the time rate of change of R_n is large. For short periods of less than an hour for which no reliable data were available, the missing values of energy fluxes were interpolated from the preceding and subsequent values.

In addition, some simple plausibility tests on the calculated latent heat flux were effective in removing spurious data associated with settling periods after instrumental servicing or with one or more of the several modes of instrumental failure described above. Hence, data with calculated latent heat fluxes greater than 400 W m^{-2} were also considered invalid, as were data for which the latent heat flux was negative when the relative humidity was less than or equal to 80%.

2.4 Fetch Analysis

An estimate was made of the surface area contributing to the micrometeorological flux measurements using the approach of Schuepp and Desjardins (Schuepp *et al.*, 1990; Desjardins *et al.*, 1992), applied with aerodynamic parameters and instrument heights appropriate for the study site. Figure 2(a) gives the result of an example calculation relevant to the eddy covariance system for a wind speed of 3.5 m s^{-1} (at a height of 6.4 m), a friction velocity of $0.38 \text{ (m s}^{-1})^2$ and a sensible heat flux of 300 W m^{-2} which shows that 80% of the flux originates within 282 m of the micrometeorological tower, with the maximum contribution at 50 m and areas 1 km away making marginal contributions. The calculated fetch has some sensitivity to the prescribed wind speed and sensible heat for which calculation is made, but this is not significant in this experimental context.

Estimates of flux contributing surface area are more complex for the Bowen ratio instrument, due to measurements at multiple heights. Figure 2(b) shows flux contributing surface area estimates for the Bowen ratio sensors mounted at heights of 3 m and 10 m. For the 3 m height, 80% of the flux measured originates within 67 m of the tower, with the maximum contribution at 20 m, while at a height of 10 m, 80% of the flux measured originates within 544 m of the tower, with a maximum contribution at 80 m. Since the vegetation within about 1 km of the tower is relatively evenly distributed, the differences in the fetch between the two Bowen ratio measurement points is not significant.

The above calculations suggest that differences in surface vegetation cover are not an issue, because the site has a fairly uniform mix of Sonoran Desert vegetation extending at least one, and more typically several kilometers, in all directions. However, available water from precipitation is an important aspect of the surface energy balance in this environment, as the summer Arizona

monsoon season is commonly associated with highly localized, convective precipitation. For example, exploratory measurements with simple non-recording rain gages deployed on a 10-m grid over a 50 m by 40 m area at this site showed that the rainfall measured in individual storms could vary by as much as 20% for convective storms and 5% for frontal storms. Thus, it is possible to receive a certain amount of precipitation within the source area from which the measured fluxes originate, but to measure a significantly different amount of precipitation in the rain gage at the study site and *vice versa*. Clearly this possibility needs to be remembered, because in principle, it has the potential to compromise the match between the measured surface energy and water budgets, at least in the short term, and so could provide a limitation on the capability to validate model calculations of surface exchanges based on single point measurement against measured fluxes from a sampled area some distance upwind.

3. OBSERVATIONAL RESULTS

3.1 *Climate Characteristics*

Measurements of the near-surface weather variables provided by the AWS confirm that the climate at the study site exhibits characteristics typical of the semi-arid Sonoran Desert climate. Figure 3 shows daily average weather variables for the year starting May 12, 1993. Daily average solar radiation ranges from 350 W m^{-2} in summer to 150 W m^{-2} in winter, with net radiation always significantly lower, because the largely cloudless skies ensure significant energy loss as long-wave radiation. Daily average air temperature varies from around 30°C in July to about 10°C in January, but the diurnal cycle in air temperature is always high, and maximum daytime temperatures in the summer commonly reach 45°C .

Precipitation is largely convective in nature, with short storms concentrated in the summer monsoon season – mainly in July and August – but the frequency of monsoon rains was lower than usual in the year for which data are available. Frontal storms contribute smaller amounts of precipitation, primarily in the winter months from December to February. The total measured precipitation at the study site over the year for which results are reported was 275 mm. Specific humidity was normally very low, about 4 g/kg for daily averages, but increases to 9 g/kg during the July–September monsoon season. Wind speeds are also fairly low, averaging about 5 m s^{-1} , again with some increase associated with storms.

3.2 *Flux Comparisons*

In principle, three independent measurements of surface energy fluxes were made simultaneously at the study site using the eddy covariance, Bowen ratio and sigma-T methods. Unfortunately, when these data were subsequently analyzed, and a careful quality control applied to each, the time period for which simultaneous data were available from all three systems was very limited. In particular, data collection with the eddy covariance system was only made early in the study year and only during the period when there were successive instrumental failures in the Bowen

ratio measurement system, i.e., prior to August 10, 1993, when routine data collection from the Bowen ratio system was finally established.

Consequently, it is only possible to present worthwhile comparisons between the surface fluxes measured with the eddy covariance method and the sigma-T method for the early period of data collection and comparisons between the Bowen ratio method and the sigma-T method for the longer, later period. Figure 4 shows comparisons between the measured sensible heat given by the thermocouple-based sigma-T method and those from the eddy covariance system in Figure 4(a) and the Bowen ratio method in Figure 4(b). Although there is a reasonable level of agreement between the measured fluxes in these two figures, there is substantial scatter and, in the case of the comparison with the eddy covariance data, evidence of a significant systematic difference at lower values of sensible heat. To isolate the cause of this systematic difference, sigma-T was recomputed using a linear average of sonic temperature, rather than a running mean of thermocouple temperature. The sensible heat flux based on eddy covariance and that based on the standard deviation of sonic temperature in Figure 4(c) compares favorably (97% correlation). This implies that the agreement between sigma-T and eddy covariance sensible heat fluxes is better when it is possible to calculate standard deviation with a linear average.

The Bowen ratio systems' decreased reliability in detecting small latent heat fluxes is important and relevant to the BATS model validation studies described later. Further, in such low flux conditions, the quality control restriction applied within these data, namely that the measured difference in atmospheric humidity between the two air intakes should be within instrumental accuracy, introduces a bias towards allowing only the highest values of latent heat in the accepted subsample. For this reason, our eddy covariance flux measurements are likely to be the more reliable of the two primary latent heat measurements made in this study, but only for the limited period for which they are available. Flux measurements from the Bowen ratio system, which were available for a much greater fraction of the time, are most credible for periods when latent heat fluxes are reasonably large. In the comparisons between modeled and measured fluxes made later, comparisons are made against data taken with both the eddy covariance and the Bowen ratio systems, but the eddy covariance data are given greater credence when the latent heat is low.

3.3 Eddy Covariance Measurements

The ability to close the energy balance is a measure of the reliability of the eddy covariance measurements. Figure 5 shows a comparison between the hourly average sum of the latent and sensible heat fluxes and net radiation minus soil heat flux. A 96% correlation between energy received and imparted by the land surface shows that an approximate energy balance is achieved, with the sum of latent and sensible heat fluxes ($L_e + H$) being, on the average, 22 W m^{-2} larger than ($R_n - G$), the measured available energy.

The diurnal pattern of eddy covariance energy fluxes are shown in Figure 6(a) for a typical dry day (July 22, 1993) with clear skies, and in Figure 6(b) for a day with substantially more cloud cover and a late afternoon storm (August 7, 1993). Typically, during dry periods, sensible heat rises

to a peak in the middle of the afternoon, while ground heat flux peaks in the morning and then decreases to reach a minimum value just after dusk. The latent heat rises during the day, but remains very low, reflecting the lack of available water. The behavior on the day with cloud cover and rain is markedly different, with substantially less net radiation and sensible heat flux, but much higher values of latent heat flux after the afternoon storm and well into the night, supported by negative ground and sensible heat fluxes.

Figure 7(a) shows hourly averaged, storage-corrected carbon dioxide flux measured with the eddy covariance system for the period between July 19 and August 9, 1993. A 5-hour running mean has also been applied to these data to improve the signal-to-noise ratio; these data are otherwise very "noisy" (Note: by comparing hourly fluctuations with this 5-hour running mean, we estimate random errors of order $0.25 \mu\text{mol m}^{-2} \text{s}^{-1}$ for hourly average measurement, but anticipate that some of this error will average out in longer-term total fluxes). For the period when carbon exchange measurements were available, carbon dioxide fluxes were always very low, averaging only about $0.25 \mu\text{mol m}^{-2} \text{s}^{-1}$ for the drier period (July 19-31, 1993) and about $1 \mu\text{mol m}^{-2} \text{s}^{-1}$ in wetter conditions (August 3 to 9, 1993). Figure 7(b) shows the diurnal CO_2 flux pattern from a typical dry day (July 22, 1993) and a typical wet day (August 7, 1993), while Figure 7(c) shows a comparison between this same wet day and the CO_2 uptake pattern for an unexceptional day during the First ISLSCP Field Experiment (FIFE, August 10, 1987; Kim and Verma, 1990). Table 1 summarizes the meteorological conditions for the wet and dry days referred to in Figures 6 and 7(b) and (c). The measured diurnal cycle of carbon dioxide exchange was about ten times larger at the FIFE site: very interestingly, the carbon uptake by the plants at this semi-arid study site continues through the night, while the prairie grasses at the FIFE site show a more typical daily cycle, with downward fluxes during the day and upward fluxes at night. We return to this interesting result in Section 5.

4. MODELING RESULTS

4.1 BATS Model Description

The Biosphere-Atmosphere Transfer Scheme (BATS) is a parameterization of the current understanding of ecohydrological processes at the scale of individual (50-1000 m) plots of vegetation. The processes incorporated are those associated with the exchange of solar and long-wave radiation, water input as rain, snow, and dew, water loss as runoff, and the surface transfer of momentum and sensible and latent heat exchanges.

The BATS uses separate model components in its radiative and hydrological description. Three layers of soil are used to calculate the water budget, all having a top surface at the soil-air interface, but with the lower surface at increasing depths. The soil is considered as a 0.1 m surface layer, a 0.5 to 2-m root layer with a total depth of 3 m. Soil temperature calculations are based on the "force-restore" method (Deardorff, 1978; Dickinson *et al.*, 1993), in which two soil layers are considered, with the upper 20-cm layer affected by the diurnal cycle and heat flux from deep soil tending to "restore" the temperature of that surface layer.

The parameterization of the vegetation canopy is based on the "single big leaf" concept, the processes considered being sensible and latent heat exchanges with the atmosphere, absorption of solar radiation and shading of the ground, and the presence of surface moisture on the canopy as a result of dew or rainfall. Excess moisture on leaves drips to the ground as throughfall. The vapor pressure over the dry area of the canopy is parameterized using stomatal resistance, which controls the escape of water from saturated conditions within the leaf to the adjacent external air. The *BATS* considers four layers of canopy in order to calculate the dependence of stomatal resistance on solar radiation. The transpiring part of plants is specified using leaf area index (*LAI*), while non-transpiring parts and dead matter are described with stem area index (*SAI*). In general, seasonal variation in *LAI* is determined by a quadric function of deep soil temperature, while the *SAI* is constant for each cover type.

The *BATS* allows for partial wetting of the canopy by rainfall, with transpiration suppressed over the wet portion of the canopy. Snow is also intercepted by leaves, and the treatment of solid water storage on the canopy is the same as for liquid water. The water stored per unit area of land surface is calculated from the difference between precipitation and evaporation from the plant surface. Two transfer phases are considered in the movement of water and heat flux from the canopy to the atmosphere: first, exchange with the air within the canopy, and second, exchange between the canopy air and the atmosphere overlying the canopy. There is no description of heat or moisture storage in the canopy air, and photosynthetic and respiratory energy transformation in the canopy are neglected. Separate temperature, humidity, and wind speed calculations are made for the canopy layer, and heat and moisture fluxes exchanged between leaves and atmosphere are transferred through this canopy layer.

The *BATS*' treatment of the Earth's surface includes bare soil, soil with vegetation cover, snow-covered soil, frozen soil, inland water, sea, and sea ice. Within each land grid square, three types of surface conditions can exist, namely bare soil, vegetated soil, and snow. In this study, the land surface without snow cover is the focus of attention. The location of canopy and bare soil is not specified within an individual grid square, and soil properties and overlying atmosphere are both assumed uniform across the square. Soil textural properties are also assumed constant with depth.

Each grid square modeled by the *BATS* is described by one of 18 land cover classes, namely mixed crop, irrigated crop, short grass, long grass, four types of forest, two types of shrubs, mixed woodland, two types of desert, tundra, glacier, marsh, ocean, and inland water. Specifying one of these cover types then selects a "standard" set of values of the surface parameters, which are then used in running the model. The *BATS* also allows specification of one of 12 standard soil texture classes which determine the hydraulic and thermal properties of the soil and eight soil color classes which determine the soil albedo.

In this study, the initial model validation run used parameters appropriate for the southwestern USA within the National Center for Atmospheric Research's (NCAR) Community Climate Model version 2 (*CCM2*), specifically standard parameters for vegetation type 11, i.e.,

"semi-desert" (see Dickinson, 1993, Table 1), soil color index 2, and soil texture class 3. To investigate whether these standard CCM2 parameters adequately characterize exchanges for the Sonoran Desert, field measurements of some of the vegetation and soil parameters were made. These measurements, together with an optimization of certain key vegetation parameters, then formed the basis of a modified set of parameters.

4.2 BATS Model Forcing Data and Initialization

The continuous set of near-surface weather variables from the AWS were expressed as 20-minute averages for the year from 12 May 1993, and were used as "forcing variables" for the BATS validation studies carried out with a stand-alone version of the BATS model, i.e., they were used as the time series of input variables required by the model to make calculations of surface flux exchange. The measured incoming short-wave radiation, air temperature and specific humidity, wind speed, and precipitation are variables used directly by the model, but the required value of downward long-wave radiation was not available from direct measurement. It was therefore derived for each time period from the surface radiation balance using measured incoming short-wave radiation and measured net radiation, together with the model-calculated surface temperature from the previous modeled time period. [Note: For this reason, model-calculated estimates of net radiation are necessarily always close to measured values in this study]. The model also requires atmospheric pressure, though in fact the sensitivity of calculated surface fluxes to variations in air pressure is small. Because routine atmospheric pressure measurements were not made, air pressure was assumed constant at 91.29 kPa, this being the average air pressure measured at nearby Tucson International Airport, with a hydrostatic correction made for the elevation difference.

In order to run the BATS model, it is necessary to designate certain initial conditions, specifically the temperatures of the canopy and soil layers and the moisture stored on the canopy and in the soil. Canopy and soil temperatures change rapidly in the model, and their initiation is therefore less critical. The initial values were specified as equal to the most appropriate measured air and soil temperatures. Unfortunately, no soil moisture measurements were available for the beginning of the BATS model validation run (on May 12, 1993). However, precipitation and temperature records suggest that soil moisture conditions for that date were similar to those for September 23, 1993 when measurements were available (both days fall at the end of extended dry periods with similar atmospheric demand). The volumetric water content of 0.136 m³ of water per m³ of soil measured on September 23, 1993 was therefore used to initialize the soil water storage in all three soil layers in the BATS model runs. In fact, the BATS was always run with 20 years "spin-up", i.e., recycling the one-year forcing data set 20 times to allow the model sufficient time to equilibrate, so the particular values used for the initiation are not critical.

4.3 BATS Parameterization

In the model calibration, a stand-alone version of the BATS was run with two sets of parameters. The first set consisted of the standard vegetation and soil-related parameters assigned to the BATS when it was used to describe the semi-arid southwestern USA in the CCM2. In the

second set of parameters a set of site- and location-specific parameters was determined in part from on-site measurements and in part by optimization, as described below. These two sets of parameters are given in Tables 2 and 3; vegetation-related parameters are given in Table 2 and soil-related parameters in Table 3.

The fractional vegetation cover at the study site, which is estimated as 40% (see Section 2.1), is significantly higher than the 10% cover fraction specified for "semi-desert", i.e., specified for the BAT5 land cover type 11, in the CCM2. Vegetation cover fraction was therefore set to a fixed value of 40% in the site-specific parameter set. No direct leaf area index (LAI) measurements were made at the site. Nonetheless, it is clear that the standard specification used in the CCM2, namely a seasonal variation between leaf area indices of 0.5 and 6.0, is unrealistic. In the absence of actual measurements, but on the advice of an ecologist with local expertise (Weltz, 1995), LAI was prescribed to have a fixed value of 1.0 in the site-specific parameter set. In practice, as we will show later, the precise value of LAI has little impact on modeled surface energy fluxes.

The fraction of roots in the upper soil layer is also a morphological aspect of semi-arid vegetation which is implausible in the CCM2 parameter set. In the CCM2, it is assumed that 80% of transpiration is extracted from the upper soil layer, despite the fact that this layer is only 10% of the overall rooting depth. Typically, roots are much more evenly distributed in Sonoran Desert vegetation (Weltz, 1995) because they need to acquire water from soil storage at depth. Accordingly, the fractional extraction from the upper soil layer was reduced to 30% which corresponds to the optimized value of the rooting fraction from the sensitivity study (discussed in Section 4.5 below), and a more even distribution of roots throughout the rooting zone (see Figure 10(c) and Table 2). The estimated vegetation height at the field site was 1.2 m (see Section 2.1). In the site-specific parameter set, zero-plane displacement height and roughness length were specified as 75% and 10% of this estimated vegetation height, i.e., $d = 0.9$ m and $z_0 = 0.12$ m, respectively. Again, we will show later that surface fluxes calculated by the BAT5 are in fact not sensitive to the precise value of these two parameters.

Soil albedo measured at the field site for near-surface soil samples is reasonably constant (see Section 2.1). Suitable values of albedo, biased towards the results for the bare soil and vegetated sites which are most representative of average conditions, are 0.20 and 0.10 for dry soil and wet soil, respectively. These values closely correspond with the BAT5 soil color index 3 (Dickinson *et al.*, 1993; Table 3), and were therefore used for the site-specific soil parameters. In practice, color index 2, which is that used in the CCM2 (see Table 3), is also in good agreement with the measured albedo at the site.

On the basis of the observed soil porosity (see Section 2.1), the BAT5 soil texture class 9 is the most representative of the soil at the field site. This soil class has a significantly higher porosity and clay content and has an order of magnitude lower hydraulic conductivity than a texture class 3 soil, the latter being the class assumed for the "semi-desert" land cover of the southwestern USA in the standard CCM2 parameter set. Thus, soil parameters in the modified BAT5 parameter set were initially chosen as class 9, but some parameters were later modified as explained below.

In early validations runs, the calculated latent heat flux when water was readily available in the soil after rain was low compared with observations, and the rate at which the calculated latent fluxes subsequently declined as soil water availability fell was much too rapid. The non-field specified parameters responsible for controlling the behavior of vegetation are minimum stomatal resistance, r_{smin} , and the two parameters (s and B) which control the "wilting" behavior of vegetation cover in response to drying soils. Because these parameters are not amenable to field measurement on a whole-canopy, area-average basis, their preferred values were determined by optimizing the correspondence between modeled and observed surface fluxes. However, in specifying the required optimum values, it was considered realistic and reasonable to place constraints on the numerical values allowed. In the case of s_w and B , the selection was restricted only to pairs of values which corresponded to the BATS soil texture classes.

The optimum value of r_{smin} so determined was 6 s m^{-1} , which is considerably less than the standard value of 200 s m^{-1} used in the CCM2. In practice, the effective value of stomatal resistance calculated and applied by the BATS during a model run is always substantially greater than this minimum value due to the assumed effect of stress factors linked to environmental (mainly meteorological) variables. In the semi-arid environment the air is usually hot and dry, which enhances the calculated effect of such environmental stress within the model. It seems that a comparatively low optimum value of minimum stomatal resistance is required to accommodate these strong environmental stress factors in the semi-arid environment, so that the calculated latent heat fluxes can agree with the transpiration rate observed when vegetation has ready access to water in the soil.

Within the BATS, the effect of soil drying on the maximum transpiration flux, E_{tmax} , is described by a "plant wilting factor", W_{LT} , through the equation:

$$E_{tmax} = \gamma_o \sum_i R_a (1 - W_{LT}^i) \quad (6)$$

where γ_o is the maximum total transpiration that can be sustained, with a summation over contributing soil layers each denoted by i , where R_a is the fraction of roots in a given soil layer. W_{LT} is zero at saturation and unity at permanent wilting point, and is calculated by:

$$W_{LT}^i = \frac{(s_i)^{-B} - 1}{(s_w)^{-B} - 1} \quad (7)$$

where s_i is the ratio of actual soil water present to that at saturation for the i^{th} soil layer, and s_w is the ratio of soil water for which transpiration essentially ceases to that at saturation in the i^{th} soil layer. B is the Clapp and Homburger (1978) exponent for the specified soil class. In this way, the two parameters s_w and B (which are linked) control the onset of plant wilting through the above equations.

In the 12 BATS soil-related parameter sets, the values of s_w and B are explicitly tied to the

values of other soil-related parameters through the assumption that there is a maximum soil water tension (of 15 mbar) against which plants can obtain water. Plant wilting behavior is thus formulated as a soil property, with a fixed relation to other soil properties (such as hydraulic conductivity); and in this way, a set of soil classes is determined by the porosity of the soil. On this basis, the soil at the field site was assigned to be the BATS soil texture class 9, as described above.

During the optimization to determine the plant wilting parameters, the other porosity-determined soil parameters such as hydraulic conductivity were held fixed, *i.e.*, they were defined to correspond to those of the BATS soil texture class 9 as described above. However, the two parameters controlling wilting behavior (s_w and B) were allowed to take on values corresponding to any of the other allowed BATS soil texture classes. The resulting optimum values of these two parameters correspond to soil texture class 4, *i.e.*, a soil class intermediate to that expected on the basis of observed soil porosity, and that (appropriate to a fairly sandy soil) assumed in the CCM2 parameter set. It is not clear whether the need to redefine these plant wilting parameters is linked to the possibility that semi-arid vegetation has evolved the ability to extract water against higher soil water tensions than other plants or whether it is of more complex reasons, perhaps because a significant proportion of the plants at the site use the Crassulacean Acid Metabolism (CAM) mechanism for carbon dioxide assimilation (see Section 5).

4.4 Model Comparisons with Observations

Hourly average latent heat flux observed using the Bowen ratio system for the period between August 26 and September 9, 1993 (day of year 238-252) is compared with that calculated by the BATS using the parameters used in the CCM2 in Figure 8(a) and using site-specific parameters in Figure 8(b). These figures illustrate observed and modeled behavior during a dry-down period after a monsoon rain in late August and are typical of the observed behavior in response to rain throughout the year. The observed latent heat fluxes show an immediate increase after each rain event (on days 238-242), followed by a steady decline during the dry-down (on days 243-252). The site-specific set of parameters captures this behavior very accurately (see Figure 8(b)), while the parameter set used in the CCM2 results in an over-rapid decline of latent heat flux after the rainy period and also does not capture the daily latent heat cycle, except for days where peak values exceed 200 W m^{-2} .

The modeled surface energy balance is significantly improved over that calculated with the parameter set used in the CCM2, which is a general result. Figure 9 shows the correlations between modeled and observed hourly average latent and sensible heat fluxes over the entire period for which Bowen ratio data are available (August 10, 1993 to March 26, 1994) and, in the case of the site-specific parameters, also shows a comparison of fluxes with measurements made with the eddy covariance system for the period between July 19 and August 9, 1993. The correlation coefficients relevant to each comparison shown in Figure 9 are higher when the site-specific parameters are used; except for Bowen ratio sensible heat fluxes where the correlation coefficients are essentially the same when the CCM2 or the modified parameter set is used (see Figures 9(b) and (d)).

In comparison with the Bowen ratio observations, the latent heat flux is generally

underestimated relative to measurements when the CCM2 parameters are used (see Figure 9(a)), but there is no significant bias with the site specific parameters (see Figure 9(c)). In the case of sensible heat flux shown in Figures 9(b) and 9(d), using the CCM2 parameters tends to overestimate large fluxes, while the site-specific parameters do not. As we observed in Section 3.2, fluxes measured with the eddy covariance system are arguably more reliable than those made with the Bowen ratio system especially when latent heat is low. Figures 9(e) and 9(f) show that calculated sensible and latent fluxes made with site-specific parameters compare well with measurements from the eddy covariance system within experimental error. In the case of observations made with the Bowen ratio system, the root mean square errors between modeled and observed sensible heat fluxes are 42 W m^{-2} and 39 W m^{-2} for the CCM2 and site-specific parameter sets, respectively; while the equivalent root mean square errors for the latent heat fluxes are 39 W m^{-2} and 24 W m^{-2} . For the eddy covariance data and the site-specific parameters, the root mean square errors between the modeled and observed sensible and latent heat fluxes are 50 W m^{-2} and 16 W m^{-2} , respectively.

4.5 Sensitivity Checks on Modified Parameters

Because the values of several parameters were redefined in part from on-site measurements and in part by optimization in the site-specific data set, it is valuable to investigate the sensitivity of modeled latent heat fluxes to these parameters.

The BATS parameters modified in response to on-site observations were fractional vegetation cover, leaf area index, the proportion of roots in the upper soil layer, zero-plane displacement, and roughness length. Figures 10(a) to 10(f) sequentially show the variation in the root mean square error and the correlation coefficient between measured and modeled latent heat fluxes calculated over the complete observational data set for a range of parameter values. In each case, the full circle shown in each of these figures is the preferred value of the parameter assigned through this study. In all cases, the root mean square error and correlation coefficient are close to their optimum with the field-specified parameter value and, in the cases of leaf area index, zero-plane displacement, roughness length, and BATS soil texture class (wilting parameters fixed to correspond to soil texture class 4) show limited parameter sensitivity.

Minimum stomatal resistance, r_{min} , and the two parameters (s_v and B) which control the behavior of vegetation cover response to drying soils merit particular attention, since they were selected not on the basis of field knowledge, but rather by parameter optimization based on the observed surface fluxes. Figures 11(a) and 11(b) show the variation in the root mean square error and the correlation coefficient between measured and modeled latent heat fluxes calculated over the complete observational data set in these two cases. They demonstrate that the preferred selections are indeed the values with minimum root mean square error and maximum correlation, but also show that selecting a values of r_{min} in the range 4 to 10 s m^{-1} and values of s_v and B for soil classes 2 to 6 do not have a major impact on the quality of fit.

4.6 Seasonal Water Balance

Table 4 shows the components of the annual water balance calculated from the BATS run using the CCM2 and site-specific parameter sets for the year from 12 May 1993. With the CCM2 parameters, BATS calculates an unreasonably large drainage to groundwater (31 mm); with the site-specific parameters, drainage to groundwater is zero, and the runoff ratio is around 5%, which is known to be a plausible value for local runoff in this region.

Table 5 gives the monthly budgets of water balance components. With site-specific parameters, the BATS is able to enhance evaporation following precipitation, and capture the observed gradual decline during subsequent dry-down. This was associated with a lower root-zone soil moisture content and zero baseflow throughout the year. Although the annual total surface runoff for both CCM2 and site-specific parameters is essentially the same (Table 4), the monthly patterns are different, with more surface runoff during the summer monsoon season and less runoff resulting from winter and spring storms for the run with site-specific parameters. In summary, the modified parameters promote the efficiency of soil water use by plants and soil evaporation, and retard the gravitational drainage out of the model soil column.

5. DISCUSSION AND CONCLUSIONS

This study, in common with other studies of surface energy balance in the southwestern USA (e.g., Kustas *et al.*, 1989; Kustas *et al.*, 1991; Nichols, 1992), confirms the fact that the energy available for return to the atmosphere largely leaves in the form of sensible heat. When viewed at both the daily and seasonal time scales, the magnitude of any latent heat flux is related strongly to the current or recent occurrence of precipitation and is less dependent on atmospheric demand. Precipitation is the most important forcing mechanism in semi-arid environments; it controls both evaporation and carbon exchange and therefore it is essential to accurately measure and characterize the distributed precipitation field to assure an accurate water balance and provide a realistic model simulation of the surface exchanges. Fulfilling this need is complicated by the large spatial variability of precipitation associated with convective storms, which are the primary mechanisms responsible for generating precipitation in semi-arid regions.

The measurements of carbon dioxide uptake made in this study are exploratory. The restricted period for which data are available correspond to a growth phase in the semi-arid vegetation, and the data do indeed show a net assimilation of carbon as expected, albeit with uptake rates much less than those commonly observed during the growth cycle of other crops in temperate regions. Arguably, the most interesting feature of these data is that they indicate that carbon dioxide uptake continues (at a reduced rate) through the night, something rarely if ever reported in other field studies. Usually a daily cycle is observed which involves uptake during the day and release at night, as in the case of the FIFE study (Verma *et al.*, 1989).

The C3 and C4 plants found at the FIFE site use a carbon dioxide assimilation mechanism which involves CO₂ entering the leaf through the open stomata during the day, with the CO₂ first converted to dicarbonic acid (malate) inside the vacuoles, followed by rapid decarboxylation of the acid to carbohydrates in the chloroplasts through photosynthesis. Though a substantial proportion

of the vegetation cover present in the Sonoran Desert (and at the field site) are C3 and C4 plants, a significant percentage of the desert vegetation consists of succulents, i.e., cacti (family *cactaceae*), agaves (*agavaceae*), and members of the *euphorbiaceae* family. The photosynthetic mechanism employed by these plants, the Crassulacean Acid Metabolism (CAM), is a two-step process (Larcher, 1994). Succulent plants open their stomata only at nighttime and assimilate CO₂ by first fixing it in the form of dicarbonic acid in the vacuoles, with the concentrations of this acid increasing throughout the night. The stomata are then closed at the onset of daylight to retard transpiration and desiccation of the plant, while the conversion of the stored acids to carbohydrates proceeds by photosynthesis. In the special case of "deciduous" plants such as the ocotillo (*Fouquieria splendens*), this process of is further complicated, because this plant employs normal C3 pathway of CO₂ fixation. However, when the soil moisture drops, the leaves of the ocotillo are shed to reduce transpiration losses to a minimum, but the plant continues photosynthesis via the stems near the leaf axles. If a sufficient fraction of the plants at the study site are of the succulent type, it is entirely plausible that there should be a net uptake of carbon dioxide sustained throughout the night during a period of growth in desert vegetation.

Micrometeorological measurements are feasible in the semi-arid conditions of this study, but it proved difficult to achieve consistent reliability and accuracy under exacting field conditions which involve extremely high temperatures and low relative humidity. The sigma-T measurements were of little use in this study, because they were of marginal and unpredictable accuracy and are, in any case, only valid for unstable atmospheric conditions - which excludes their use between dusk and dawn and during and immediately after rain. After some substantial initial problems, it was ultimately proven easier to obtain a reasonably reliable record of surface energy fluxes using the Bowen ratio method rather than the eddy covariance method, primarily because using the Bowen ratio method required much less time for data analysis, less computational resources, and substantially less in-field maintenance and calibration. Nonetheless, we believe measurements made with the eddy covariance system are likely to have greater accuracy and are therefore preferable when greater accuracy is required over short measurement periods.

Using the set of vegetation and soil related parameters used in the CCM2, the BATS model was unable to capture the observed rapid increase in the outgoing latent heat in response to precipitation and the gradual falloff during subsequent dry-down. Using these parameters gave a poor description of daily, day-to-day, and seasonal behavior of the surface energy balance. In contrast, an excellent description was achieved using a modified, site-specific set of parameters which was based on a few simple observations and some optimization of key vegetation-related parameters. This modified set of parameters more realistically reflected the fact that there is a sparse but fairly constant vegetation cover of around 40% at the study site and that the soil contains a significant amount of clay. Further, these parameters suggest that the vegetation responds quickly when it receives water, with a substantial and rapid increase in photosynthesis and transpiration. Ultimately the soil dries and the plants reach a wilting point which, in the BATS model, is assumed to be determined by soil porosity. In practice the desert plants then seem able to sustain transpiration by accessing water from soil in the root zone longer than might be expected on the basis of the soil's high porosity alone. The uniqueness of the photosynthetic processes employed by many desert plants

and the effectiveness of the transpiration process revealed by this exploratory study argue for further measurement and modeling studies in this unusual and interesting environment.

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Table 1. Meteorological conditions at the study site for the typical dry and wet days shown in Figures 6(a), 6(b), 7(b) and 7(c).

Meteorological Variables	Typical Dry Day (July 22, 1993)	Typical Wet Day (August 7, 1993)
Min. Air Temperature (°C)	24.9	19.5
Max. Air Temperature (°C)	34.9	33.0
Min. Soil Temperature (°C)	29.2	25.1
Max. Soil Temperature (°C)	45.1	40.3
Min. Relative Humidity (%)	16.4	35.9
Max. Relative Humidity (%)	44.5	84.1
Min. Wind Speed (m s ⁻¹)	0.20	0.75
Max. Wind Speed (m s ⁻¹)	4.5	9.9
Total Daily Precipitation (mm)	0.0	17.5

Table 2. The BATS vegetation-related parameters as used in the National Center for Atmospheric Research Community Climate Model version 2 (CCM2) for "semi-desert" (BATS land cover class type 11) and the modified site-specific parameters which produced improved model description of the observed data.

Parameter Name	Standard CCM2 Parameters	Improved, Site-Specific Parameters
Maximum fractional vegetation cover	0.1	0.4
Seasonality factor (<i>i.e.</i> difference between maximum fractional vegetation cover and fractional cover at a temperature of 269 K)	0.1	0.0
Roughness length (m)	0.1	0.12
Zero-plane displacement height (m)	0.0	0.9
Minimum stomatal resistance ($s\ m^{-1}$)	200.0	6.0
Maximum leaf area index	6.0	1.0
Minimum leaf area index	0.5	1.0
Stem and dead matter area index	2.0	2.0
Inverse square root of leaf dimension ($m^{-1/2}$)	5.0	5.0
Light sensitivity factor ($m^2\ W^{-1}$)	0.02	0.02
Vegetation albedo for $\lambda < 0.7\ \mu m$	0.17	0.17
Vegetation albedo for $\lambda \geq 0.7\ \mu m$	0.34	0.34
Depth of upper soil layer (m)	0.1	0.1
Depth of rooting zone soil layer (m)	1.0	1.0
Depth of total soil layer (m)	3.0	3.0
Fraction of water extracted by upper layer roots (saturated)	0.8	0.3

Table 3. The BATS soil-related parameters as used in the National Center for Atmospheric Research Community Climate Model version 2 (CCM2) for the "semi-desert" areas of southwest USA (BATS soil texture class 3) and the modified site-specific parameters which produced an improved model description of the observed data.

Soil Parameter	Standard CCM2 Parameters (soil texture class 3)	Soil Texture Class 9	Modified Tucson Parameters (modified from soil texture class 9)
Porosity (volume of voids to volume of soil)	0.39	0.57	0.57
Minimum soil suction (mm)	30.0	200.0	200.0
Saturated hydraulic conductivity (mm s^{-1})	0.032	0.0022	0.0022
Moisture content relative to saturation at which transpiration ceases	0.151	0.455	0.266
Exponent "B" defined in Clapp & Hornberger (1978)	4.5	8.4	5.0
Ratio of saturated thermal conductivity to that of loam	1.3	0.85	0.85
Soil color index	2	3	3

Table 4. Components of the annual water balance and net radiation for the year from 12 May 1993, as calculated with the *BATS* using measured weather variables. Two model parameter sets are used: (1) the National Center for Atmospheric Research Community Climate Model version 2 (*CCM2*) parameters for the "semi-desert" areas of the southwestern USA, and (2) a set of modified site-specific parameters which produced an improved model description of the observed data.

Component	<i>BATS</i> Model Using Standard <i>CCM2</i> Parameters	<i>BATS</i> Model Using Modified Tucson Parameters
Net Radiation (mm)	1181	1180
Precipitation (mm)	275	275
Evaporation (mm)	230	262
Surface Runoff (mm)	14	14
Baseflow (mm)	31	0

Table 5. Monthly water balance and root-zone soil moisture content (*RSW*) as calculated with the *BATS* using measured weather variables. Values shown are for the set of modified site-specific parameters; values in parentheses are for *CCM2* parameters for the "semi-desert" areas of the southwestern USA.

Month	Precipitation (mm/month)	Evapo- transpiration (mm/month)	Surface Runoff (mm/month)	Baseflow (mm/month)	Monthly Mean <i>RSW</i> (mm)
January	0.3	11.6 (6.5)	0.0 (0.0)	0.0 (2.6)	137.7 (162.9)
February	35.7	18.8 (19.4)	1.2 (2.2)	0.0 (2.8)	149.9 (172.7)
March	30.6	28.6 (26.3)	1.1 (0.9)	0.0 (3.1)	150.8 (171.1)
April	3.2	21.8 (11.4)	0.0 (0.0)	0.0 (2.5)	138.0 (163.0)
May	14.9	14.9 (16.5)	0.1 (0.2)	0.0 (2.4)	101.8 (125.2)
June	0.1	1.2 (8.9)	0.0 (0.0)	0.0 (2.0)	129.8 (158.1)
July	14.7	14.3 (16.9)	0.1 (0.2)	0.0 (1.9)	133.4 (159.3)
August	100.9	36.4 (46.7)	8.2 (6.9)	0.0 (2.3)	148.3 (170.0)
September	15.0	51.7 (27.8)	1.1 (0.5)	0.0 (3.1)	164.5 (178.6)
October	16.0	25.0 (17.6)	0.3 (0.3)	0.0 (2.9)	146.3 (169.0)
November	35.9	20.9 (20.7)	1.3 (2.4)	0.0 (3.0)	149.5 (171.9)
December	8.1	16.4 (11.5)	0.2 (0.1)	0.0 (3.0)	147.9 (168.3)

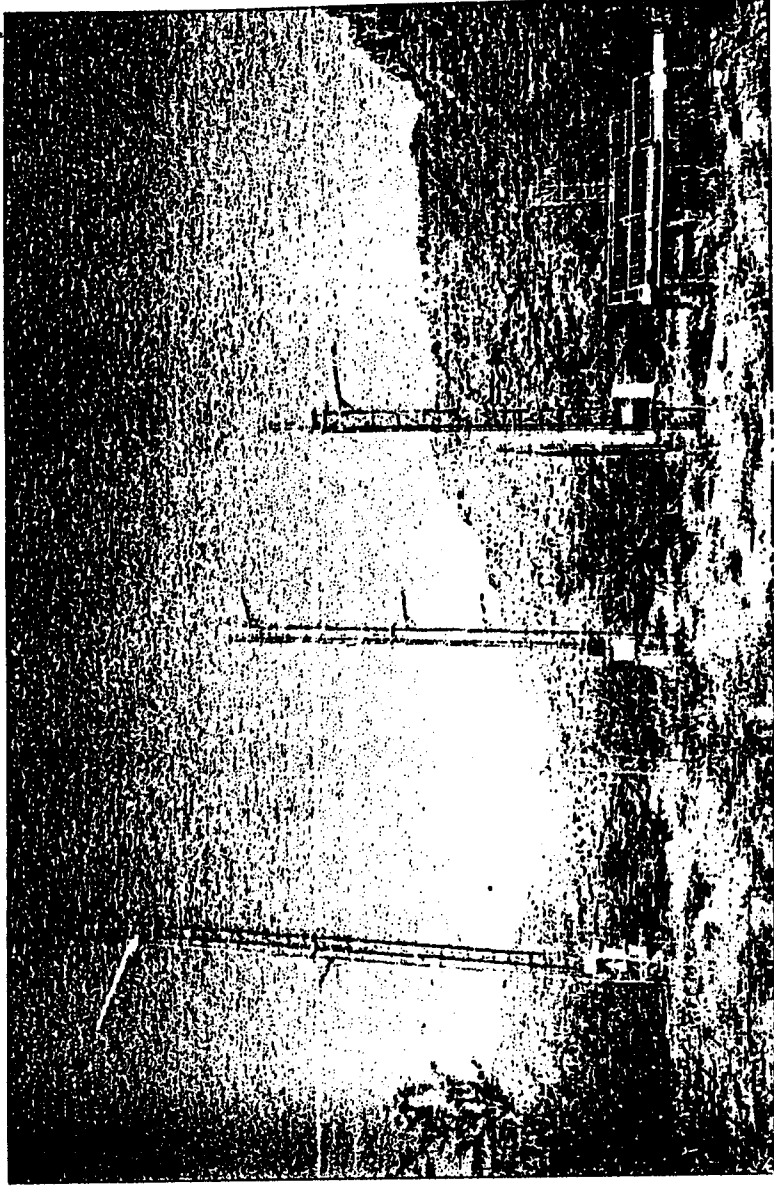


Figure 1. Micrometeorological instruments at the field site west of Tucson, Arizona, viewed from the direction of the prevailing winds and longest fetch towards the Tucson mountains, which are approximately 1 km distance towards the northeast. The Bowen ratio system is shown mounted on the tower on the left, the automatic weather station and sigma-T systems are mounted on the center tower, and the eddy covariance sensors are mounted on top of the tower on the right. The remainder of the eddy covariance system, including the photovoltaic array used to provide power, are mounted on the ground to the right.

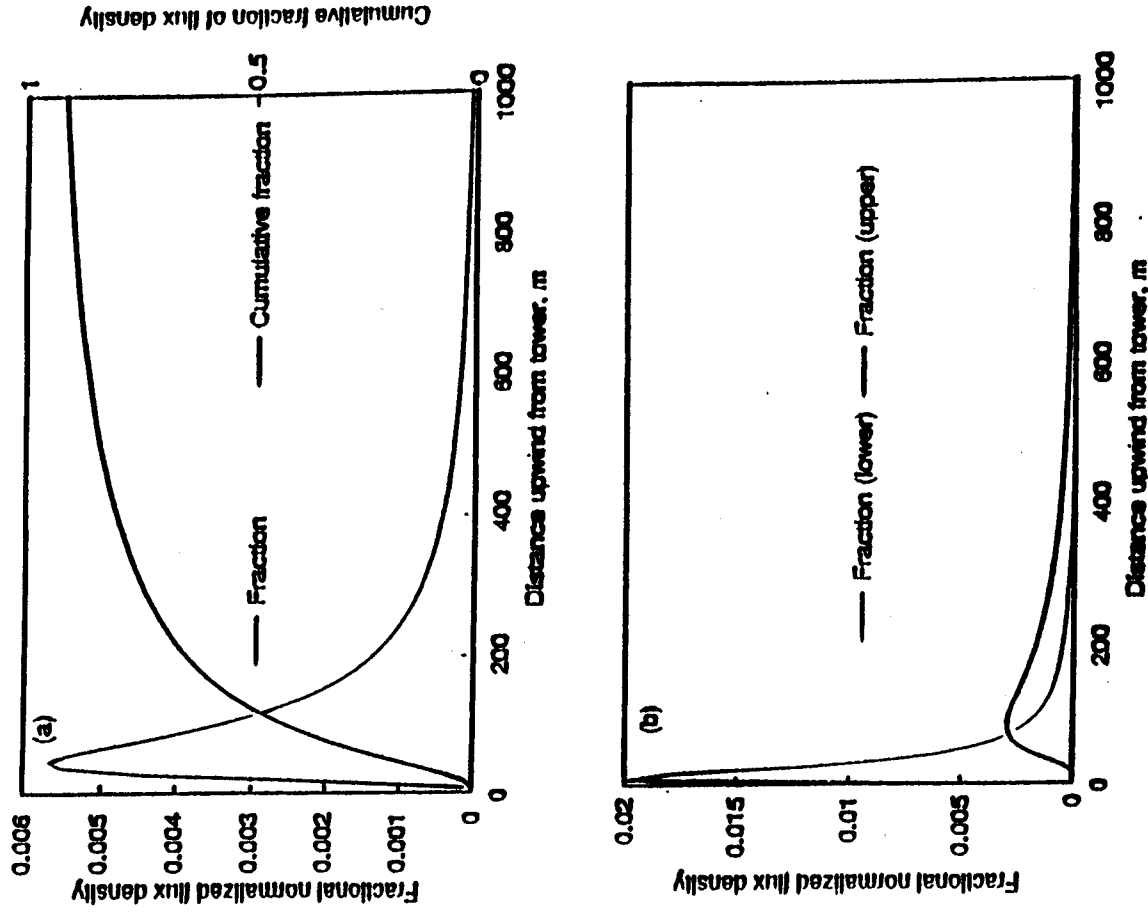


Figure 2. (a) The calculated fractional flux density versus distance upwind of the field site for eddy covariance observations made at 6.4 m based on the method of Schnepp *et al.*, (1990). The calculations assume a zero plane displacement of 0.9 m and a roughness length of 0.12 m and are made for a wind speed of 3.5 m s^{-1} , a friction velocity of $0.38 \text{ (m s}^{-1})^2$ and a sensible heat flux of 300 W m^{-2} . (b) The calculated fractional flux density versus distance upwind of the field site for Bowen ratio observations made at 3 m and 10 m.

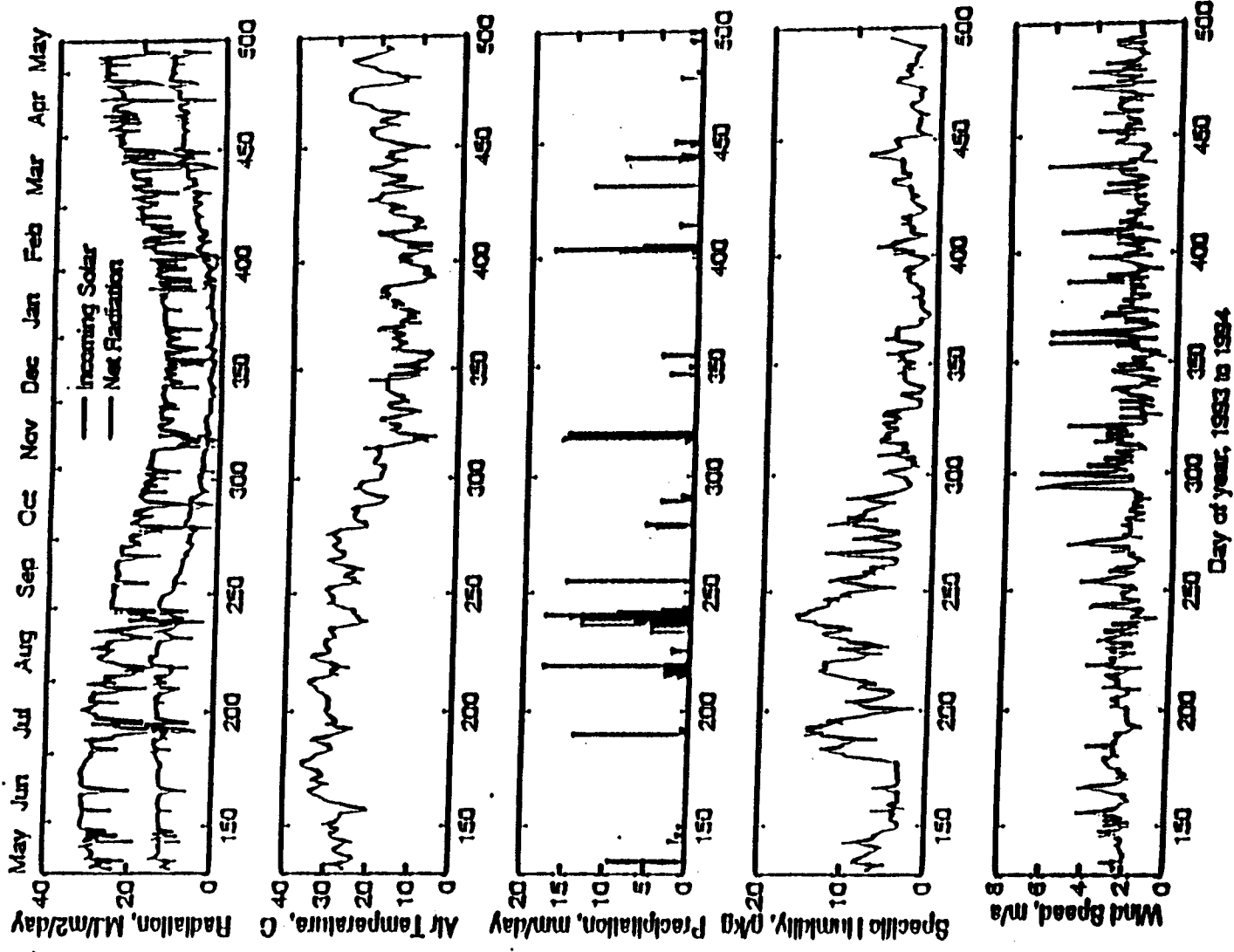


Figure 3. Daily total values of precipitation, solar and net radiation, and daily (24-hour linear average) values of air temperature, specific humidity and wind speed measured at the field site for the year from 12 May 1993.

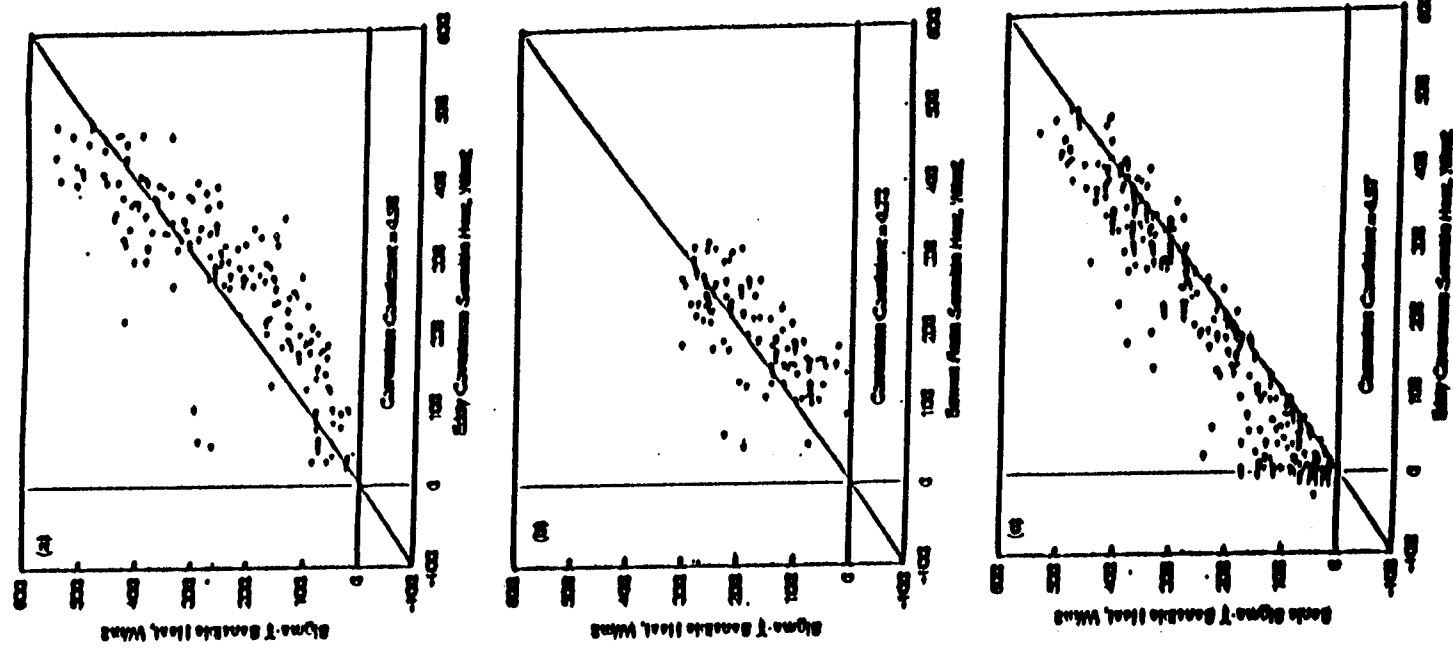


Figure 4. Measured flux comparisons of sensible heat given by the sigma-T method and those from (a) the eddy covariance system, and (b) the Bowen ratio method. (c) Measured flux comparisons of sensible heat given by the eddy covariance and sigma-T based on sonic temperature.

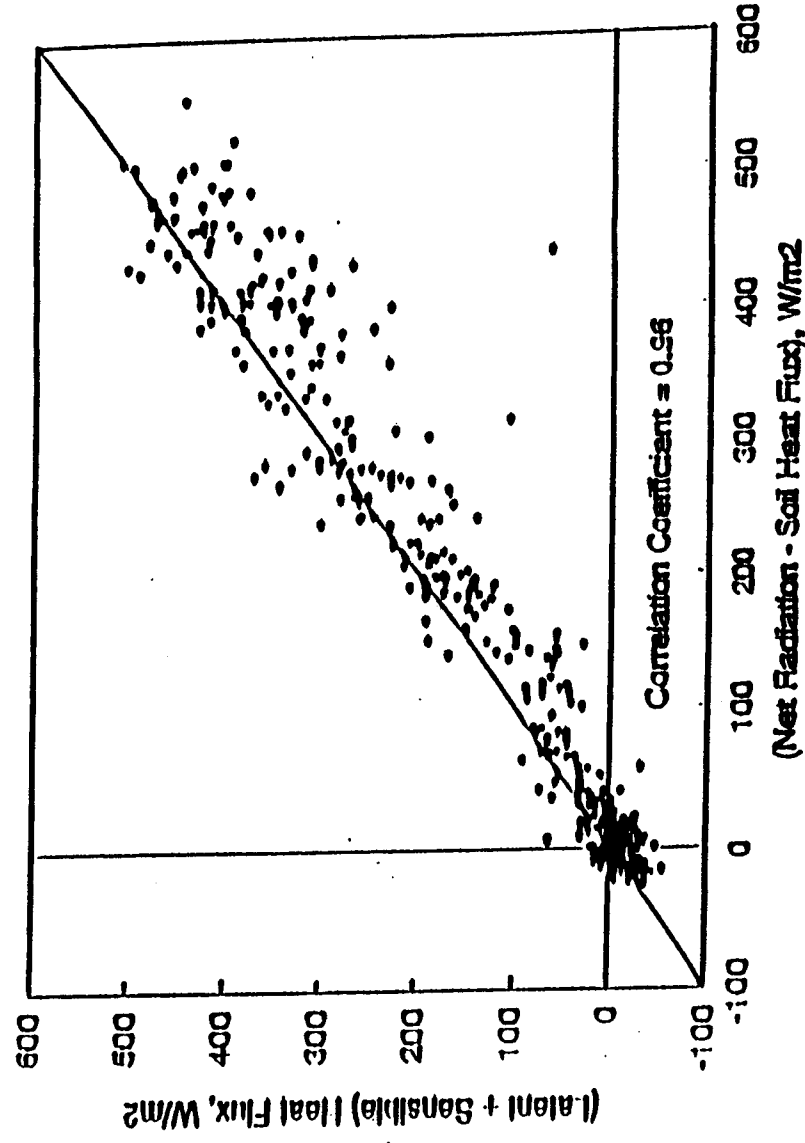


Figure 5. Comparison between the hourly average sum of the sensible and latent heat fluxes measured with the eddy covariance system and the measured net radiation minus the soil heat flux for equivalent hours.

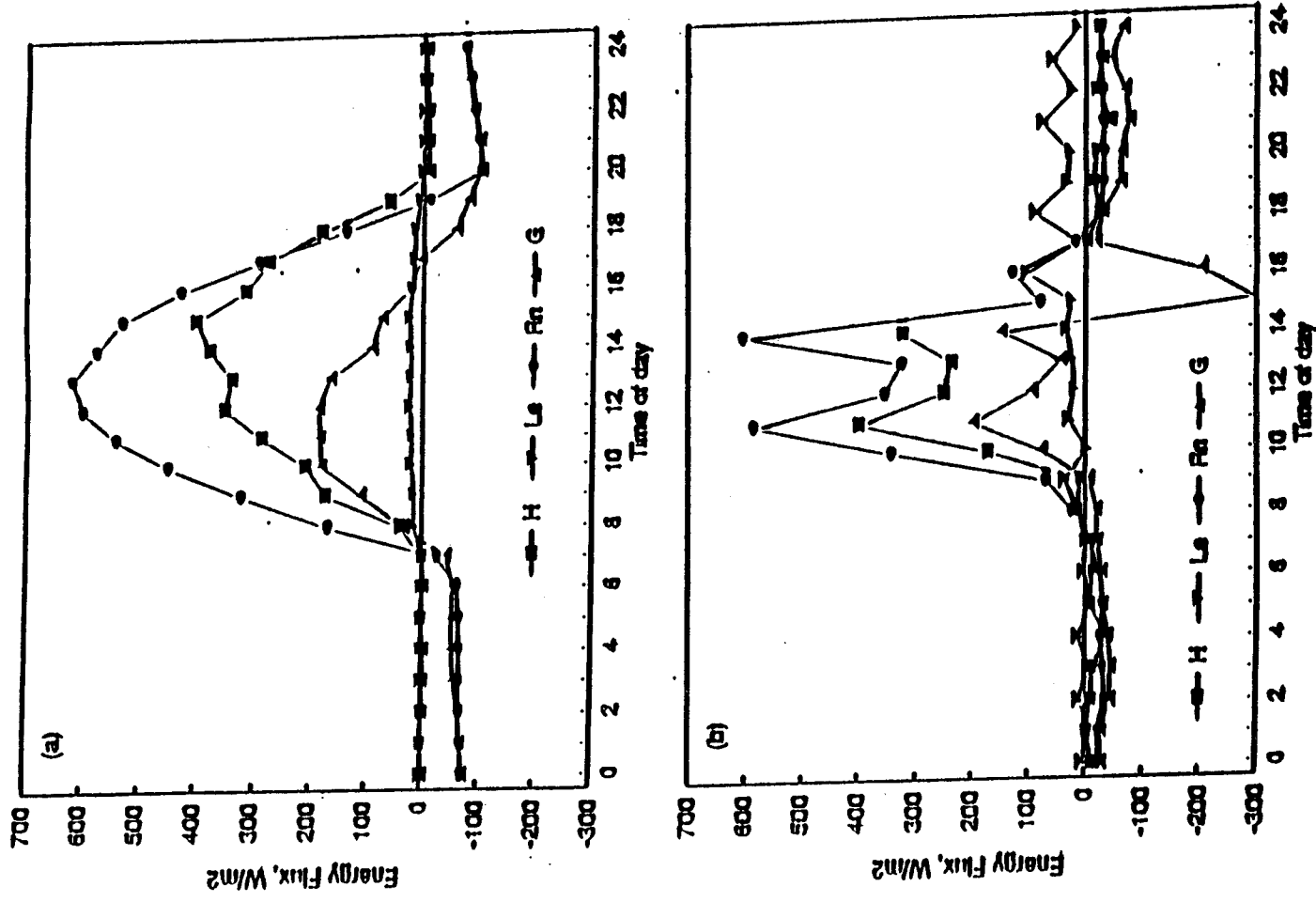


Figure 6. The diurnal pattern of surface energy fluxes measured by the eddy covariance system (a) for a typical dry day with clear skies (July 22, 1993), and (b) for a day with substantially more cloud and a late afternoon rain storm (August 7, 1993). In each diagram the sensible heat flux is labeled H , the latent heat flux L_e , the net radiation flux R_n , and the soil heat flux G . For reference, the daily solar maximum occurs at about 12:30 pm local time.

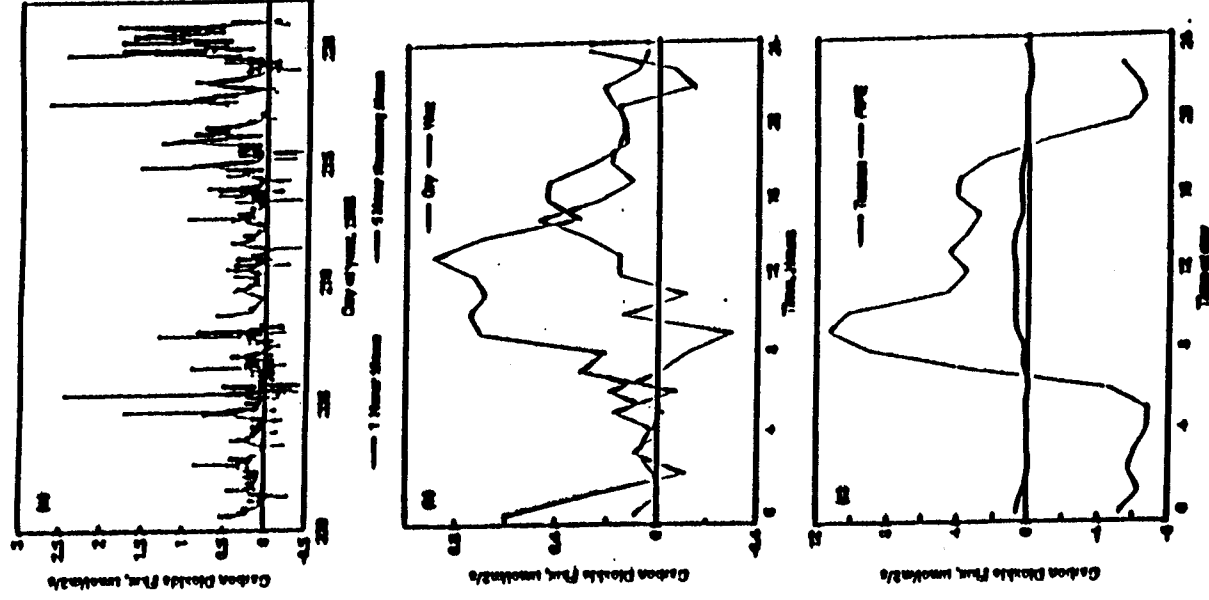


Figure 7. (a) One hour mean and 5-hour running mean of the storage-corrected carbon dioxide flux measured with the eddy covariance system for the period between July 19 and August 9, 1993, (b) the diurnal carbon dioxide flux at the experimental site on a typical dry day (July 22, 1993) and on a typical wet day (August 7, 1993), and (c) the diurnal carbon dioxide flux at the experimental site on this typical wet day in comparison with that measured during the First ISLSCP Field Experiment (FIFE, August 10, 1987; Kim and Verma, 1990). Notice that the measured fluxes for the Sonoran Desert are lower than those measured at the FIFE site and, at this time of year are generally downward, even at night.

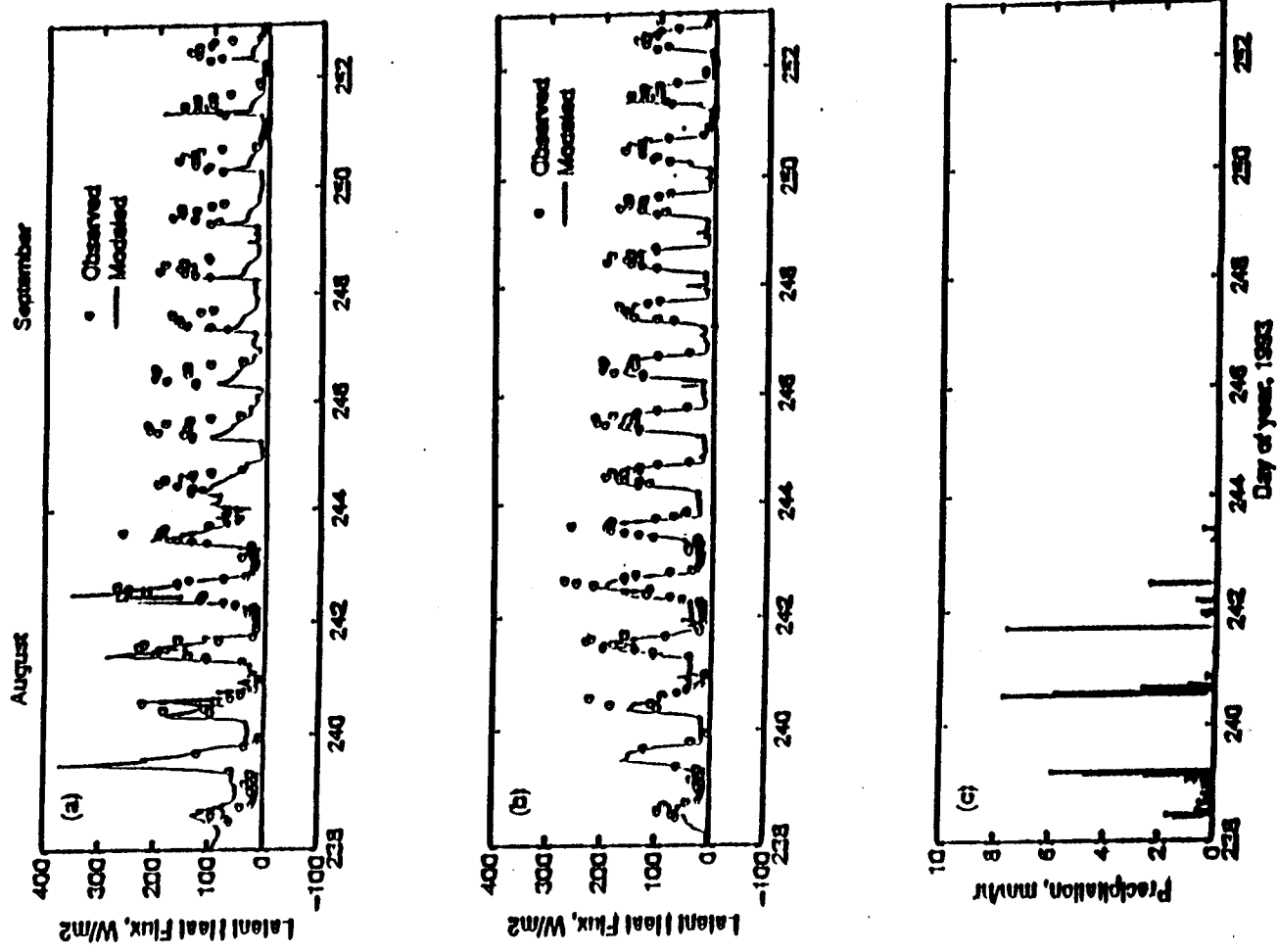


Figure 8. Hourly average latent heat flux observed using the Bowen ratio system for the period between August 26 and September 9, 1993 is compared (a) with that calculated by the *BATS* using the parameters used in the *CM2*, and (b) using site-specific parameters. Panel (c) shows the observed hourly precipitation.

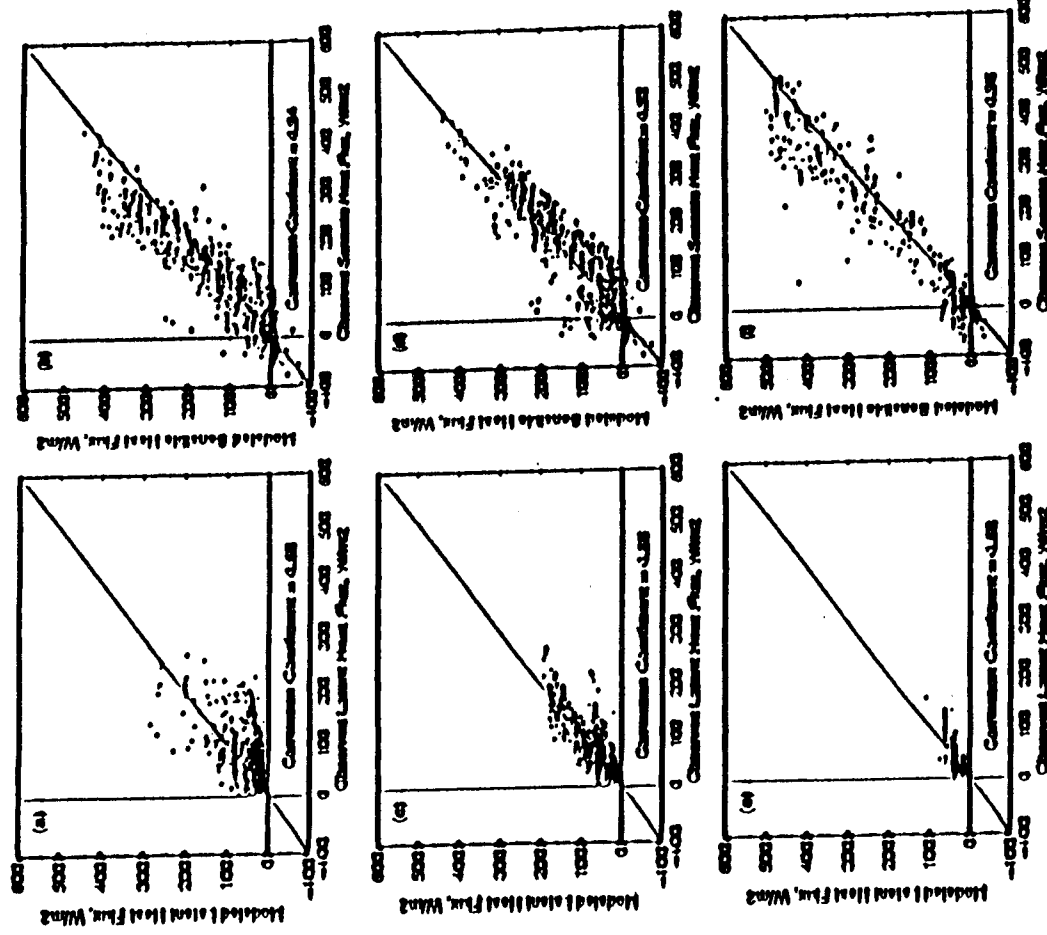


Figure 9. Correlations between modeled and hourly average observations of latent and sensible heat flux. Figures (a) and (c) show latent heat flux, and Figures (b) and (d) show sensible heat flux measured with the Bowen ratio system. Figure (e) shows latent heat flux, and Figure (f) shows sensible heat flux measured with the eddy covariance system. The modeled fluxes in Figures (a) and (b) are calculated using the parameter set used in the CCM2, while those in the remaining figures are made using the site-specific parameters defined in this study. The eddy covariance data were collected between July 19 and August 9, 1993, while the Bowen ratio data were collected between August 10, 1993 and March 26, 1994.

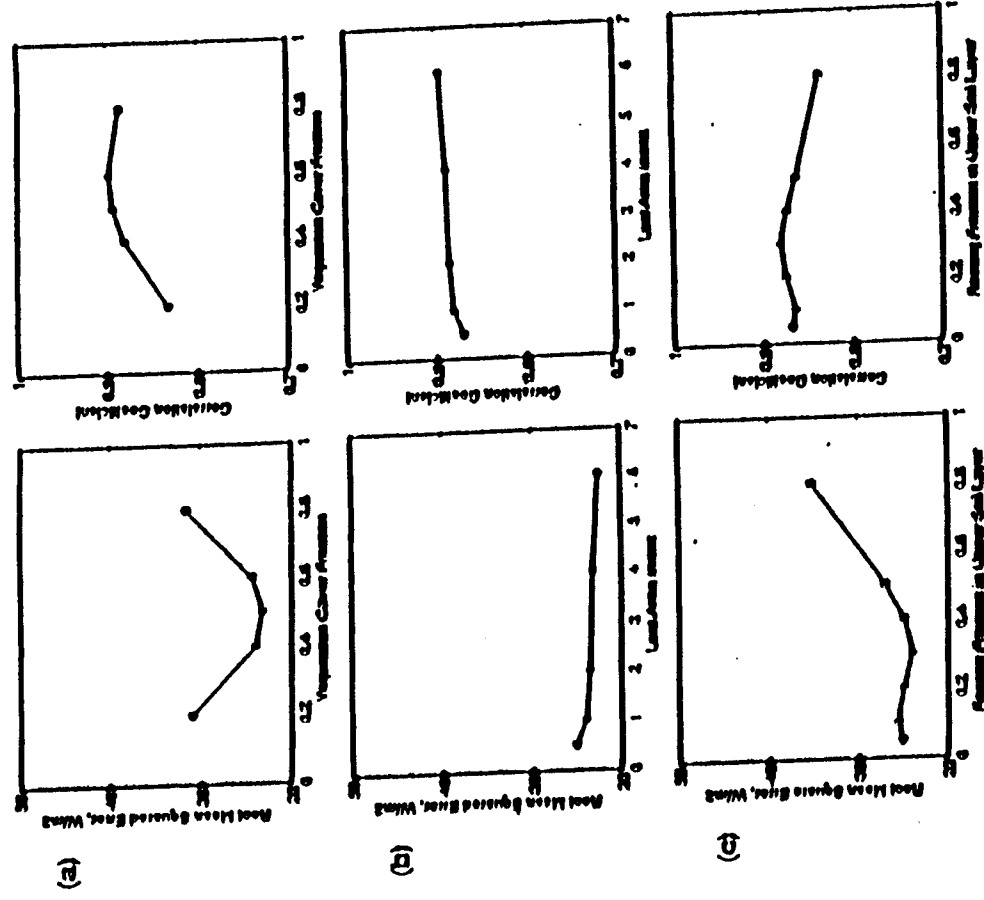


Figure 10. Variation in the root mean square error (in $W m^{-2}$) and the correlation coefficient between measured and modeled latent heat fluxes calculated over the complete (Bowen ratio) data set for a range of values of (a) fractional vegetation cover, (b) leaf area index, (c) the proportion of roots in the upper soil layer, (d) zero-plane displacement, (e) roughness length, and (f) *BATS* soil texture class for other than plant wilting parameters (wilting parameters are fixed to correspond to soil texture class 4). In each case, the full circle shown in each of these figures is the preferred value of the parameter.

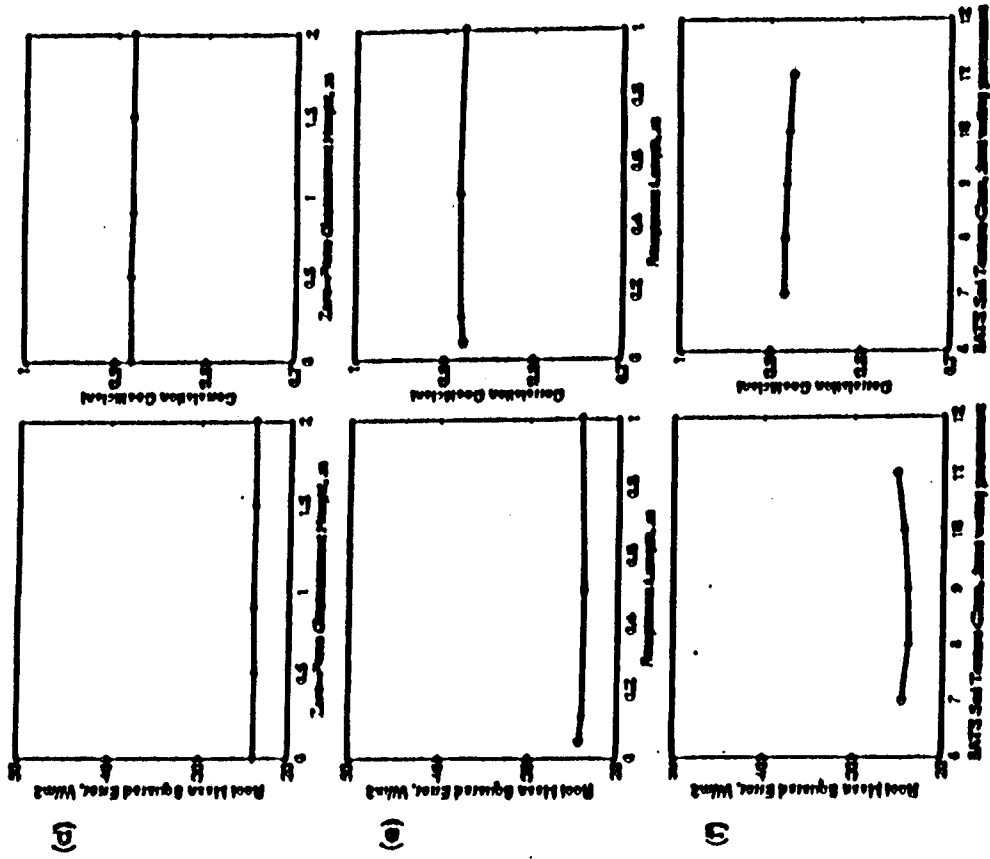


Figure 10 (continued).

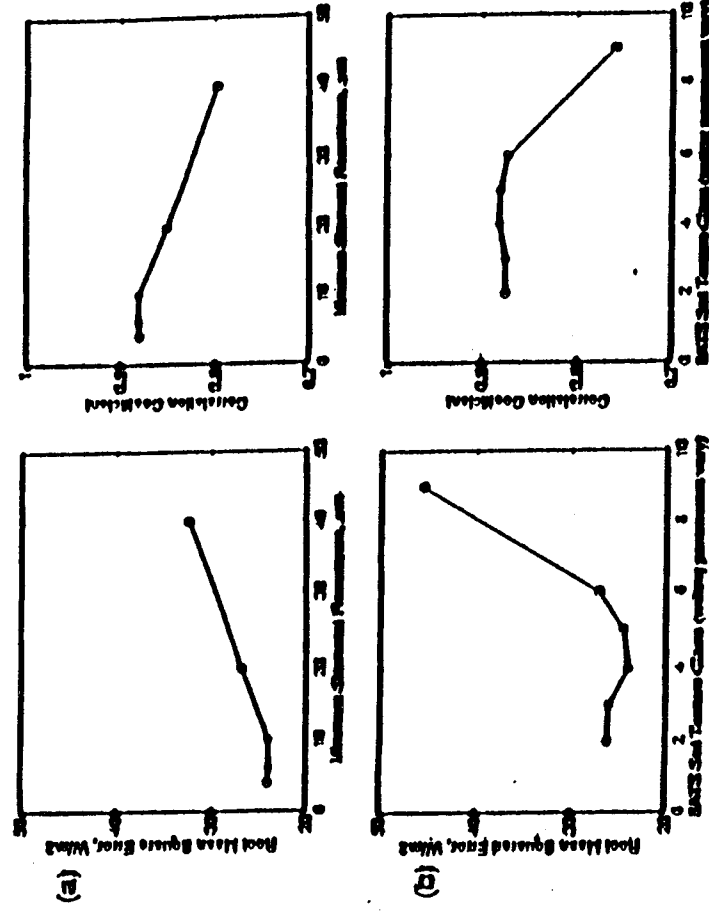


Figure 11. Variation in the root mean square error and the correlation coefficient between measured and modeled latent heat fluxes calculated over the complete observational data set for a range of values of (a) minimum stomatal resistance, and (b) the *BATS* soil texture class for plant wilting parameters (soil parameters not associated with wilting are fixed to correspond to soil texture class 9).

Research Note

**The aggregate description of semi-arid vegetation with precipitation-generated
soil moisture heterogeneity**

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Abstract

Meteorological measurements in the Walnut Gulch catchment in Arizona were used to synthesize a distributed, hourly-average time series of data across a 26.9 by 12.5 km area with a grid resolution of 480 m for a continuous 18-month period which included two seasons of monsoonal rainfall. Coupled surface-atmosphere model runs established the acceptability (for modeling purposes) of assuming uniformity in all meteorological variables other than rainfall. Rainfall was interpolated onto the grid from an array of 82 recording rain gauges. These meteorological data were used as forcing variables for an equivalent array of stand-alone Biosphere-Atmosphere Transfer Scheme (BATS) models to describe the evolution of soil moisture and surface energy fluxes in response to the prevalent, heterogeneous pattern of convective precipitation. The calculated area-average behavior was compared with that given by a single aggregate BATS simulation forced with area-average meteorological data. Heterogeneous rainfall gives rise to significant but partly compensating differences in the transpiration and the intercepted rainfall components of total evaporation during rain storms. However, the calculated area-average surface energy fluxes given by the two simulations in rain-free conditions with strong heterogeneity in soil moisture were always close to identical, a result which is independent of whether default or site-specific vegetation and soil parameters are used. Because the spatial variability in soil moisture throughout the catchment has the same order of magnitude as the amount of rain falling in a typical convective storm (commonly 10% of the vegetation's root zone saturation), in a semi-arid environment non-linearity in the relationship between transpiration and the soil moisture available to the vegetation has limited influence on area-average surface fluxes.

1. Introduction

This research note addresses the question, “In the extreme case of the semi-arid environment subject to convective rainfall, are the ensuing real patterns of soil moisture such that they have significant impact on the area-average description of vegetation in meteorological models?” It is best read as a sequel to Arain et al., (1996).

Arain et al. investigated the adequacy of simple rules for defining the effective, area-average (or aggregate) value of the vegetation-related parameters required to specify surface-atmosphere interactions for heterogeneous mixes of vegetation covers. They used data from the First International Satellite Land Surface Climatology Field Experiment (FIFE; Sellers et al., 1988) to validate and initiate the BATS-ABL coupled model of interactions between land surfaces and the atmospheric boundary layer -- see Arain et al. (1996) for details of this model. In most conditions they found that so-called aggregation rules (Shuttleworth, 1992) can provide area-average values of vegetation parameters which, when applied within the Biosphere-Atmosphere Transfer Scheme (BATS; Dickinson et al., 1986), calculate values of surface fluxes that are acceptably similar to those calculated with explicit recognition of different patches of vegetation cover in the coupled model.

As a side-issue in their paper, Arain et al. (1996) discovered that when they included a significant proportion of irrigated crops amongst heterogeneous vegetation in an otherwise dry landscape, the performance of the hypothetical aggregation rules was compromised. This result is explained by the non-linear response of vegetation to changing soil moisture in the BATS and, it is assumed, in nature as well -- see Arain et al. (1996) for greater detail.

An analysis of satellite-derived land cover classes shows that the proportion of irrigated land

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falling within an individual grid square used in meteorological models is rarely greater than 10% in the USA. (The USA is the only region for which such data sets are readily available at this time.) Moreover, coupled modeling studies using the BATS-ABL show that including irrigated vegetation with this relative extent still allows the use of aggregation rules with acceptable accuracy in the simulated area-average fluxes. Hence the presence of patches of artificially wetted (irrigated) soil is not a serious issue.

However, Arain et al. (1996) suggested that in certain circumstances extreme differences in soil moisture might occur naturally, and they recommend that a study be made to determine if poor aggregation rule performance exists in natural situations. This note explores this specific issue for the deliberately extreme case of soil moisture heterogeneity resulting from intermittent localized convective rain falling in a semi-arid region of the southwestern USA.

2. Field Data

The data used in this study is from the Walnut Gulch Experimental Watershed (31°43'N, 110°00'W), which is located 120 km southeast of Tucson, Arizona, and which comprises the upper 150 km² of the Walnut Gulch drainage basin (see Figure 1). Here soil types range from clays and silts to well-cemented boulder conglomerates (Renard et al., 1993), with gravelly and sandy loam surface textures containing an average of 30% rock and little organic matter (Kustas and Goodrich, 1994). The topography comprises gently rolling hills incised by rather steep drainage channels. The mixed grass-brush rangeland vegetation, which is typical of southeastern Arizona and southwestern New Mexico, ranges from 20 to 60% in coverage.

The catchment receives 250-500 mm of precipitation annually, two-thirds of it falling as convective precipitation during a summer monsoon season encompassing the months of July through

mid-September. The balance of precipitation mainly falls during winter frontal storms of Pacific origin. The monsoonal precipitation is usually high-intensity convective thunderstorms of limited areal extent. The average run-off in the ephemeral streams is small (~5%) and of short duration, which implies that one-dimensional models, like the BATS, can provide a reasonably realistic description of this region's surface-atmosphere interactions. The 85 recording rain gauges which are currently maintained in and around the watershed--an average of one rain gauge per 1.76 km², along with 11 runoff measuring flumes, and two sets of weather and energy flux measuring stations are illustrated in Figure 1.

2.1. Synthesis of Distributed Data Arrays

This modeling study commences on July 16, 1990 and continues through December 31, 1991. It therefore includes two seasons of monsoon rainfall, the first season being wetter than average and the second season drier than average. The modeling strategy (described below) requires a two-dimensional, distributed array of hourly-average near-surface meteorological forcing variables for the stand-alone version of the BATS for the 18-month study period. These data were synthesized from available measurements and projected on to a 56 column by 26 row array that covers the Walnut Gulch catchment at a 480 m grid resolution (Figure 1). The 651 grid squares which are located within the boundary of the Walnut Gulch catchment are those for which model comparisons were made in this study.

Comparisons between the meteorological observations at the two weather and energy flux measuring stations showed that when it was not raining, the inter-site differences in solar radiation, temperature and vapor pressure deficit are low (typically less than 20 Wm⁻², 1 °C, and 0.14 kPa respectively), even when the difference in the observed soil moisture at these sites was at a

maximum. Some larger, short-term differences were observed (particularly in temperature and vapor pressure deficit) while it was raining if the intensity and, in particular, the timing of rainfall differed between the two sites. Large short term differences in solar radiation were typical on cloudy days. However, model sensitivity tests using the two observed time series of weather variables as alternate forcings demonstrated that the overall results of the study are not sensitive to this meteorologic variability. Moreover, studies with the BATS-ABL coupled model (described later) show that area-average surface fluxes have little sensitivity to atmospheric advection that might have been generated by the most extreme observed differences in soil moisture.

In consequence of the just-described observations and modeling results, it is appropriate to assume that the values of incoming solar radiation, near-surface air temperature and vapor pressure deficit required by the stand-alone version of the BATS are, in effect, constant across the 26.9 by 12.5 km study area. For reasons of data consistency, the preferred values of meteorological forcing variables applied during modeling were those from the Kendall micrometeorological station, see Figure 1. However, for about 1% of the time data was missing from this site so data from Lucky Hills was substituted, and for a few brief periods (less than 1% of the time) when neither station was operational, it was necessary to use data from a nearby Fort Huachuca meteorological station. Tests made at times when data from both Walnut Gulch and Fort Huachuca were available confirmed that differences in these forcings were insignificant for modeling purposes.

Distributed precipitation fields were provided by interpolation from the rain gauge network. To ensure consistency, interpolation was made from the 82 gauges for which continuous rainfall data was available for the whole study period. Raw rain gauge data was digitized from analog charts and then converted into an hourly time series for the 18-month study period. Previous studies with this rain gauge network (Haider, 1994) have demonstrated that the multiquadric-biharmonic technique

is the preferred method of interpolating precipitation because it combined practicality with acceptable accuracy. An hourly time series of precipitation data was synthesized for each node in the distributed data grid by applying this technique at each hour of the study period to interpolate the rain gauge data to the equivalent grid location.

2.2. BATS Model Calibration

To test the sensitivity of the specific aggregated results to the vegetation parameters in the BATS, the aggregate modeling tests (described later) were performed using both the default BATS semi-arid vegetation parameters, and the site-specific parameter values observed at the Walnut Gulch study site.

Performance of these two parameter sets was evaluated by comparing the resulting modeled surface fluxes with the measurements at the two observation sites. However, the value of this comparison was restricted by known weaknesses in the flux measuring instruments used at these sites. The observation of sensible heat flux was made at a height of 9 m using the eddy correlation method with a single “Gill” propeller to measure the vertical wind velocity. A previous comparison with relevant energy- and water-balance data (Williams, 1996) indicates that the routine use of this wind sensor results, on average, in about a 30% under-measurement of sensible heat flux. Therefore, a fixed correction of this size was applied to the measured sensible heat flux prior to its use in the energy budget to calculate latent heat flux for the purpose of comparison with model calculations.

The standard BATS parameters appropriate to the Walnut Gulch study site corresponded to surface cover/vegetation class number 11, the semi-desert default (from the 18 available classes), soil textural class 3 (from 12 textural classes ranging from sand [1] to clay [12]), and soil color class 2 (from the 8 available color classes ranging from light [1] to dark [8]) -- see Dickinson et al. (1986)

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for definition of these classes. Simulations using these parameters provided poor agreement with observed surface energy fluxes, both in terms of the overall magnitude of the fluxes (latent heat fluxes being on average underestimated by 58%, while sensible heat fluxes were overestimated by 33%), and in terms of the modeled daily pattern (the model calculated that latent heat fluxes fell rapidly after an early morning peak while observations suggested a more uniform daily pattern).

Selection of site-specific parameters was largely based on the literature, much of which resulted from the Monsoon-90 field campaign which was held in the Walnut Gulch catchment from July 23 to August 10, 1990, and which has been extensively documented (e.g. Kustas et al., 1991, 1994a, 1994b, 1994c, Kustas and Goodrich, 1994, Stannard et al., 1994, Moran et al., 1994, Humes et al., 1994, Wertz et al., 1994, Menenti and Ritchie, 1994, Hipps et al., 1994). These sources allowed the confident specification of many BATS vegetation parameters (including percentage vegetation cover, roughness length, zero-plane displacement height, albedo, and maximum and minimum leaf area indices), and many soil parameters (including porosity, saturated hydraulic conductivity, and wet and dry soil albedos). In addition, a prior modeling study carried out at a Sonoran Desert site near Tucson, Arizona (Unland et al., 1996) suggested that two other parameters (the minimum stomatal resistance of the vegetation and the fraction of water extracted from the upper soil layer) require revision if the BATS is to provide realistic simulation of semi-arid vegetation. The values of these two parameters were therefore determined by a simple optimization process against the observed surface energy fluxes in an analogous way to Unland et al. (1996).

Use of the revised, site-specific parameters greatly enhanced the quality of the comparison between the BATS modeled fluxes and observed fluxes, resulting in correlation coefficients of 0.875, 0.912, and 0.970 and root mean squared errors of 41.7 Wm^{-2} , 50.8 Wm^{-2} , and 20.3 Wm^{-2} for latent, sensible, and ground heat flux respectively.

3. Modeling Studies

This project explored two modeling issues. The first was to check the validity of assuming uniform meteorology (apart from precipitation) across the Walnut Gulch catchment while the second appraised the impact of heterogeneous precipitation on vegetation-atmosphere interactions.

3.1. Coupled Modeling Tests

The BATS-ABL model was used to investigate the significance of possible patch-to-patch atmospheric advection in the presence of extreme soil moisture heterogeneity observed across the study site. An initial run was made using the distributed array of stand-alone BATS models forced with the distributed array of meteorological data (see below) to determine the most extreme differences in modeled soil moisture status in the catchment. The soil moisture status and soil and vegetation temperatures for the two most extreme elements in the model array were selected and used as initial values for moist and dry patches in BATS-ABL, while the area-average meteorology at the time these extreme values occurred was used to initiate the atmospheric variables in that model -- see Arain (1994) and Arain et al. (1996) for greater detail on this initiation process.

The BATS-ABL was run from these initial conditions with a modeled domain size of 2000 m and with alternate vegetation patches (with wet and dry soil) each 1000 m wide -- again, see Arain et al. (1996) for greater detail on procedures. On the basis of the surface-atmosphere interaction simulated with the BATS-ABL, area-average values of the surface exchanges and of the near-surface meteorological variables at 2 m were calculated across the modeled domain. A separate calculation of the area-average surface exchanges was then made with the area-average value of the meteorological variables at 2 m used as forcing variables (rather than the spatially varying values of forcing variables that had been modeled to occur at this height in the BATS-ABL run). In this way

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an assessment could be made of the acceptability of assuming a uniform meteorology (in all but precipitation) across the Walnut Gulch catchment in the distributed array model runs. The results of this coupled modeling study were reassuring in that the calculated area average fluxes given by the BATS-ABL (which acknowledge spatial variability in near-surface meteorology coupled to variability in soil moisture) agreed with those calculated using an enforced uniformity in near surface meteorology to within 11.5%, 8.6%, and 7.1% for latent heat, sensible heat, and ground heat flux respectively in these conditions of most extreme soil moisture difference.

3.2. Aggregation Modeling Studies

Two stand-alone BATS modeling studies were used to assess the impact of heterogeneous precipitation on vegetation-atmosphere interactions. An individual (aggregate) BATS model and a distributed array of single point BATS models were created and identically parameterized over the Walnut Gulch Watershed. The distributed BATS model was driven with the spatially distributed meteorological forcing data whose derivation is described above, while the aggregate BATS model was forced with meteorological data which was the area-average of that used for the distributed array for each hour. (Recall that, in practice, precipitation is the only variable considered to have significant spatial heterogeneity).

The calculated surface exchanges for the array of single point BATS models were averaged and compared to the output from the aggregate BATS model which, it was hypothesized, provided an adequate area-average description. In this way the area-average interaction with the realistic representation of precipitation-induced heterogeneity in soil moisture can be compared with that given when heterogeneity is neglected in the modeling process. If the two outputs disagree with one another, then failure of aggregation rules (analogous to that which Arain et al. (1996) reported as

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possible with patches of irrigated agriculture) can be deemed to have occurred in these albeit extreme, but nonetheless natural, conditions.

The results indicate no evidence of significant aggregation rule failure with naturally-occurring differences in soil moisture. Modeling results are presented here only for runs using the site-specific BATS parameters, but simulations were also made with the BATS default parameters (see above). Although the area-average surface energy fluxes calculated with the default parameterization in both the aggregate and the distributed array BATS models differed from the equivalent runs using the site-specific parameters, none of the conclusions given below are affected because these conclusions are based on the difference between aggregate and distributed array runs made with the same set of model parameters.

Figure 2 shows a comparison between the calculated latent heat, sensible heat and ground heat fluxes given by the aggregate BATS model and those given by the distributed array of BATS models for hours when there was no intercepted water on the vegetation canopy during the 18 month study. Clearly surface energy fluxes agree extremely well, to within a few $10\text{'s } \text{Wm}^{-2}$ -- this being the criterion for success adopted by Arain et al. (1996).

Notwithstanding this success in rain-free conditions, there is an interesting side-effect of spatial variability in precipitation which is apparent in this study, and which is related to the interception of rainfall by the vegetation canopy. When the rain falling in a storm covers a large proportion (approximately 60% or more) of the modeled study area there is an inconsequential difference between the calculated components of evaporation for simulations made with the aggregate BATS model and the distributed array of BATS model. However, when storms of limited extent occur, this is not the case. Figure 3 shows an example of the area-average, model-calculated evaporation flux for the study area that originate (a) as transpiration, and (b) as the evaporation of

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intercepted water, along with (c) the interpolated rainfall for 3 days when there were 7 short storms, 3 of which covered approximately 25% of the modeled area whereas the remaining 4 covered less than 10% of the modeled area. Inspection of Figure 4 reveals that localized storms of this type give rise to differences in the modeled area-average rate of evaporation of intercepted water between the aggregate and distributed array models which can be as large as 30-50% for short periods.

This difference arises because the aggregate model rainfall is (wrongly) assumed to be equally distributed across the modeled area so all elements in the grid contribute to the net interception loss, and its modeled interception rate is therefore higher. At the same time, there is (in part) a compensating difference in the area-average transpiration flux, because it is suppressed for the whole study area and not just for the rained-on portion. However, the compensation between these two components of evaporation loss is not perfect and, in general, the erroneous enhancement of calculated interception loss for the aggregate BATS model slightly exceeds the erroneous reduction in the calculated transpiration loss during the storm. Consequently, more water evaporates and less rain infiltrates during storms, causing the modeled soil moisture and transpiration to be slightly less on rain-free days. This small reduction in inter-storm, area-average transpiration is 1.8% as modeled. Table 1 lists this difference, as well as the differences between the cumulative water balance components for both simulations.

4. Discussion and Conclusions

On the basis of the above comments, it is clear that heterogeneous rainfall can result in a poor description of short-term evaporation rates during and immediately after rainfall because the description of the rainfall interception process is imperfect in a one-dimensional, aggregate representation. This feature has been recognized for some time, see, for example, Shuttleworth

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(1988). However, in the case of semi-arid regions, where there is little runoff and where water availability is the limiting control on evaporation, maintaining the water balance necessarily requires that modeled transpiration is reduced between rain storms to compensate for this additional within-storm loss. This is not necessarily the case in regions where evaporation rates are energy limited and where runoff is plentiful.

Notwithstanding this last mentioned complication, the primary focus of the present study was to investigate whether realistic patterns of soil moisture have significant impact on the area-average description of vegetation in meteorological models in the extreme case of a semi-arid environment subject to intermittent but intense convective rainfall. The results are definitive in this respect. In rain-free conditions, but with natural, rain-induced, heterogeneous soil moisture, the difference between the calculated surface energy fluxes given by a single, aggregate BATS model and that given by a distributed array of BATS models is very small indeed.

The basic reason for this is that the amount of rain in an individual precipitation event is always a fairly small proportion of the total amount of water that can be stored within the rooting zone of the simulated vegetation. In this study the area-average soil moisture stored in the rooting zone varies with season: typically the soil moisture store is about half full during the monsoon season, and close to empty after an extended dry period. However, since local convective storms occur randomly in space, spatial variations in available soil moisture across the landscape at any time have the same order of magnitude as the amount of rain in one typical rainfall event. Convective rain storms can be heavy and localized in this environment -- the largest storm observed at any grid point in the modeled array during the study period was about 50 mm, for instance, but the amount of water that can be stored in the rooting zone of the vegetation (336 mm) is large in comparison with this. Hence, at any time, only about 10% of the non-linear relationship between soil moisture and

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evaporation is being sampled within a given soil moisture pattern, and the non-linear relationship can be considered approximately linear over this small, sampled range. Given that this is the case in this extreme semi-arid situation, it is likely that it is more generally true elsewhere.

Acknowledgement

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Figure Captions

Figure 1.

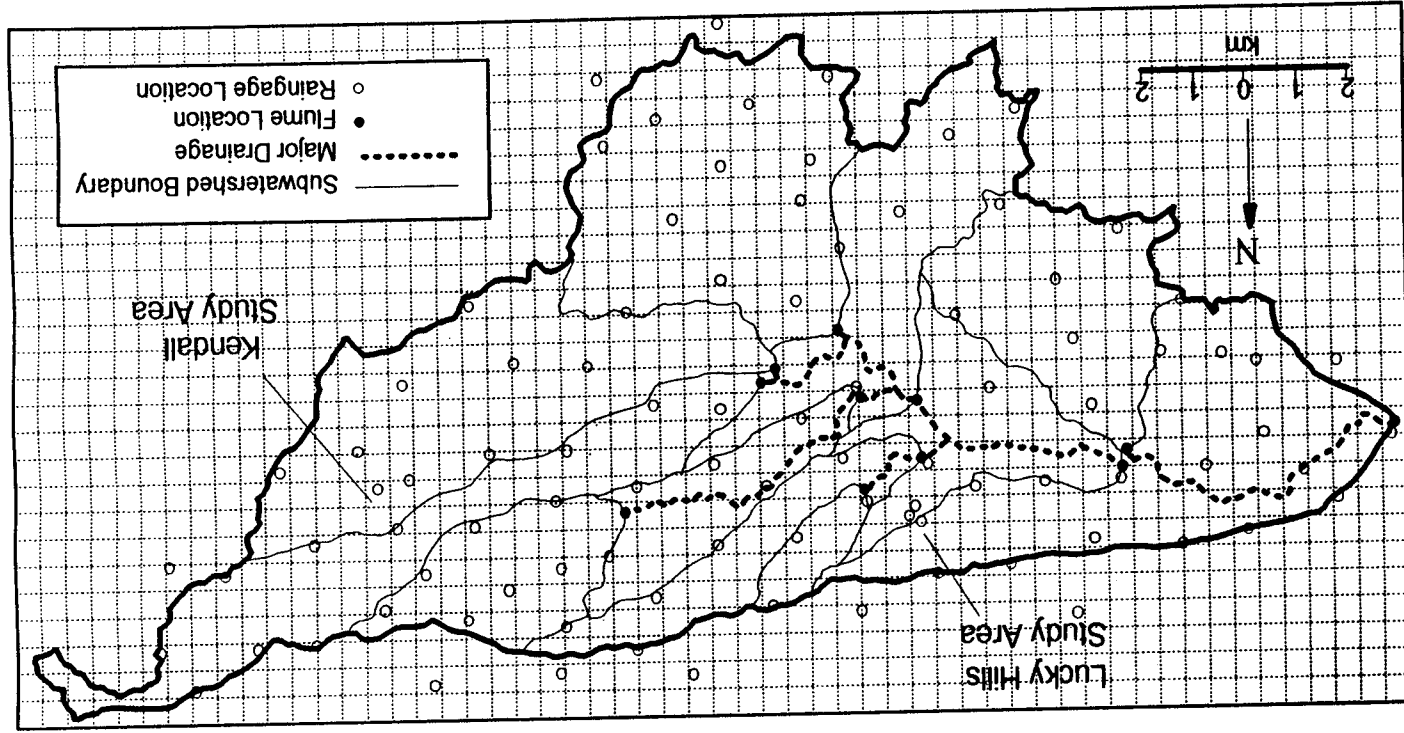
The area described by the distributed array of stand-alone BATS models which includes the Walnut Gulch catchment in Arizona. The figure shows the catchment boundary, the location of rain gauges, flumes and the Kendall and Lucky Hills micrometeorological stations.

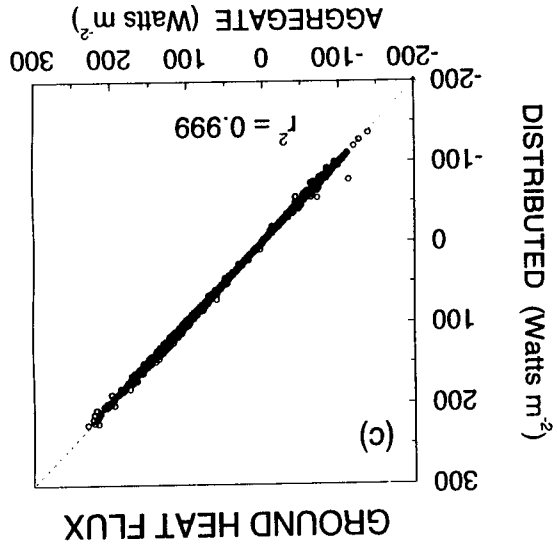
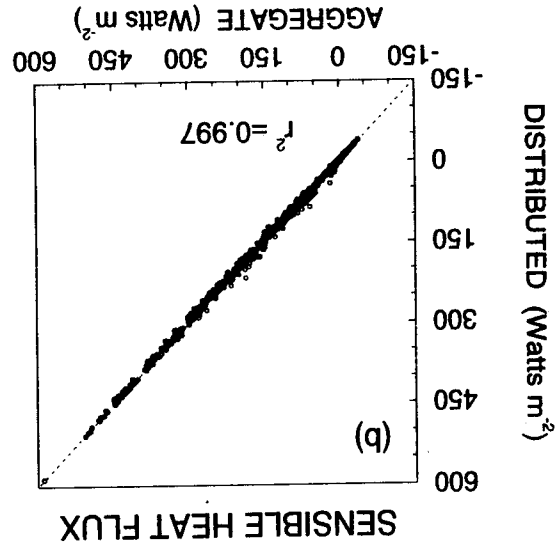
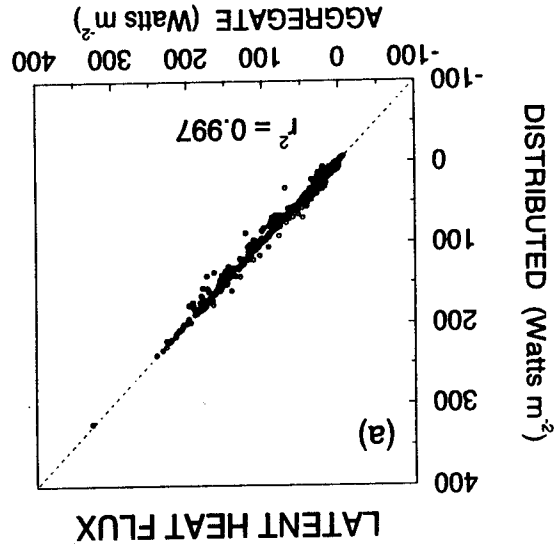
Figure 2.

Comparison between the hourly-average (a) the latent-heat flux; (b) sensible-heat flux; and (c) ground heat flux as calculated by the single, aggregate BATS model forced with area-average meteorological data and the distributed array of BATS models forced with distributed data for all rain-free hours in the 18 month study period

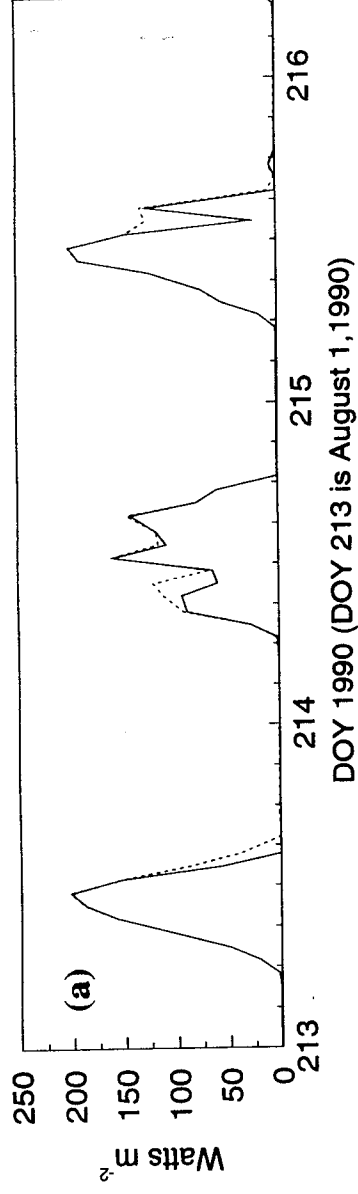
Figure 3.

Time series of calculated rates of (a) transpiration, (b) evaporation of intercepted water as given by the single, aggregate BATS model forced with area-average meteorological data (full line) and the distributed array of BATS models forced with distributed data (broken line) and (c) the interpolated precipitation from day of year 213 to 215 of 1990 (August 1- August 3, 1990). There were 7 localized rain storms during this period.

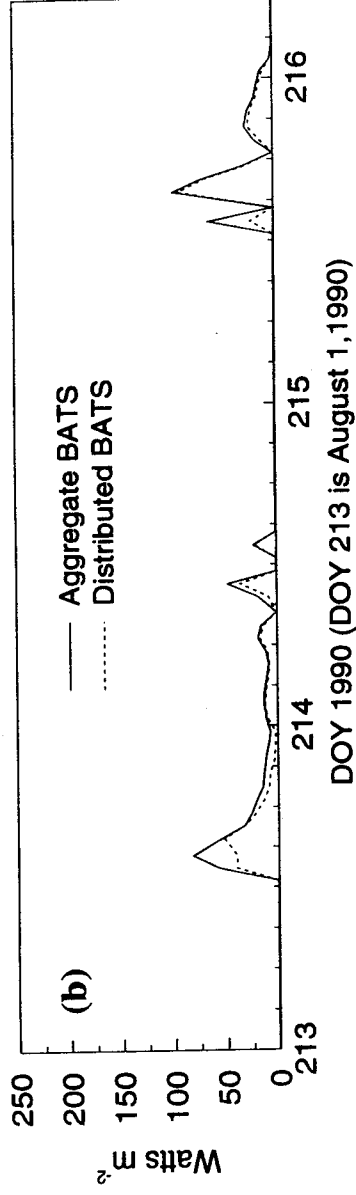




Transpiration



Evaporation of Intercepted Water



Interpolated Precipitation

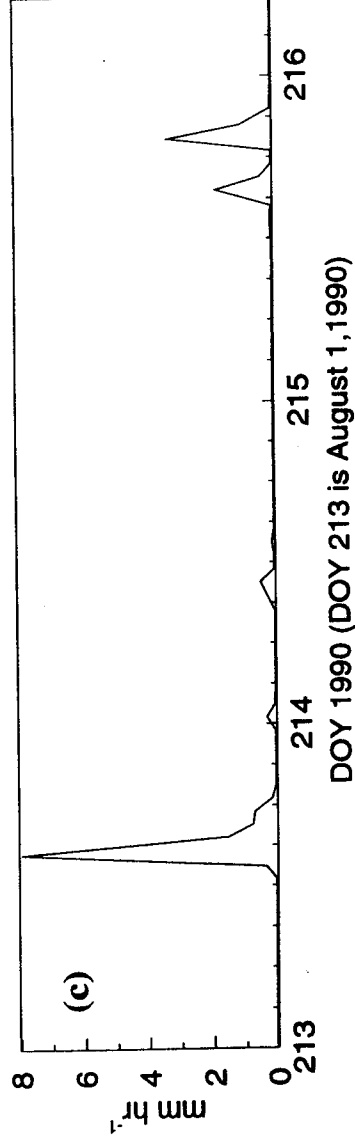


Table 1 Modeled Water Balance Components for the Aggregate and Distributed *BATS* models using site-specific parameters.

Modeled Water Balance from July 16, 1990 - December 31, 1991 (534 Day Model Run)			
Cumulative Water Balance Component	Aggregate Model (mm)	Distributed Model (mm)	% Difference *
LE Flux	466.27	467.29	0.22%
Precipitation	426.46	426.34	-0.03%
Run Off	21.33	20.33	-4.92%
ΔStorage from Soil Moisture	-54.53	-55.05	0.99%
Transpiration	359.41	366.05	1.81%
Interception	25.91	19.51	-32.80%
Soil Evaporation	73.62	74.59	1.30%
Water Balance Closure			
	Σ Input (mm)	Σ Output (mm)	Σ In / Σ Out
<i>Aggregate</i>	459.66	466.27	98.6%
<i>Distributed</i>	461.06	467.29	98.7%

* [(Distributed - Aggregate) / Distributed] x 100%

Patch scale aggregation of heterogeneous land surface cover for
mesoscale meteorological models.

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Abstract

Theoretical studies dealing with aggregation of surface parameters at small scale are reviewed. Finding effective parameters for surface resistance is possible for most cases by taking simple geometric or arithmetic averages of the component resistances. The use of more sophisticated techniques such as the blending height improve the calculations. Resistances for heat and water vapour behave differently in heterogeneous terrain. A simple surface energy balance model is adapted to show the behaviour of the roughness length of heat and water vapour in heterogeneous terrain. It is suggested that this simple parameterization can adequately take into account the effect of variation in surface cover on the fluxes of heat and water vapour.

Introduction

The parameterization of subgrid variability within large scale models of the atmosphere appears to be Achilles heel of current attempts in modelling global climatic change (e.g. Enkathabi and Eagleson, 1989). The difficulty in parameterizing subgrid variability of any kind is that information on subgrid processes is carried in the models only in an integrated, area averaged, sense. Meteorological models which operate at gridsizes larger than the scale of the land surface variability thus need to parameterize the variability in land surface cover in an efficient, yet physically realistic way. This paper is concerned with parameterizing the effect of small (patch) scale variability of land surface cover on the fluxes of heat, momentum and water vapour, but would apply also to most other scalar properties such as CO_2 .

Several studies have addressed the problem of the effect of spatial variation in surface cover on the average fluxes of heat, momentum, and water vapour. These studies can be divided into three categories: those of an experimental nature, those of a theoretical nature, and relative straightforward applications of modelling concepts in large scale models of the atmosphere. Examples of the first category dealing with surface roughness variability may be found in Mahrt and Ek (1993) and Grant (1991) amongst others. Application of various algorithms based on the simultaneous calculation of fluxes from each patch individually (e.g. Dolman, 1992) in climate models may be found in Avisar (1992) and Koster and Suarez (1992). These approaches, however, usually

neglect advective processes. This paper is primarily concerned with the theoretical approach, which tries to improve on the previously mentioned studies by taking advective processes into account.

Typical length scales of interest are of the order of a few hundreds of meters to 10 kilometres. This type of variation corresponds to type A in the distinction made by Shuttleworth (1988) and De Bruin (1989) and many others since. Type A variation refers to small, disorganized variation in land surface cover which shows only an aggregated response in the boundary layer and is not sufficiently large to create an individual response in the boundary layer or to establish mesoscale circulations. This type A landscape is typical of many agricultural areas in Europe and indeed, the analysis of HAPEX-MOBILHY (André et al, 1985) data gave rise to this distinction. Also, the soil moisture patterns, arising from localized convective activity in the semi-arid tropics, may fall into this category of variability.

This paper attempts to review the studies aimed at providing guidelines for aggregation of heterogeneous land surface cover at scales less than 10 km, i.e., scales where the effects of land surface variations blend into a homogeneous boundary layer. An example of the use of a non-hydrostatic mesoscale model (Kalthof *et al.*, 1993) is shown to investigate the effect of realistic surface variations on the aggregate representation. Simple parameterizations based on the concept of sparse crop energy balance models (Dolman, 1993a, Blyth and Dolman, 1994) are then used to show the

behaviour of the area average roughness lengths of heat and momentum for heterogeneous land surfaces.

THEORY

At small scales, say less than 10 km, it is possible to distinguish two types of heterogeneity, denoted in this paper by sparse crop heterogeneity and small scale land cover variability (Fig 1). Sparse crop heterogeneity has been subject of several modelling efforts in the last few years (e.g. Shuttleworth and Wallace, 1988, Dolman, 1993). In sparse crops surface energy balance models large areas of bare soil or undergrowth interact with a canopy at a level $z_0 + d$. These models operate in a quasi one-dimensional manner, forcing a two-dimensional system into a one-dimensional algebraical framework. The height of the crop, h , is the relevant vertical height scale and the frictional velocity, u_* , the appropriate velocity scale at canopy or surface layer level. Raupach (1992) presents a framework to estimate the relevant horizontal length scales involved in this:

$$\frac{u_* h}{u x} \quad (1)$$

where u is the horizontal windspeed and x the horizontal length scale of the variations. Using typical values of $u_*/u = 0.1$ and $h = 1$ to 10 m, suggests the horizontal size of the areas for which this approach may be valid to be around 10 to 100 m, or possibly larger in the case of forests.

It is possible to extend this sparse crop framework to larger areas, by following the same strategy of applying a one-dimensional framework to solve for a two-dimensional system in nature. Rather than the two surfaces interacting at the level $z_0 + d$, they now interact at a level higher in the atmosphere, *above the canopy*: the blending height. As in the previous case of sparse crop models, the two-dimensional interaction of the surfaces is modelled in a framework which appears to be one-dimensional: this is to make the problem analytically more tractable. The blending height is defined here as that height where it is appropriate to make the assumption that the fluxes of momentum and heat are still approximately in local equilibrium, whereas the mean properties such as windspeed and concentration are sufficiently well mixed to assume that they are (horizontally) homogeneous (Wieringa, 1986, Mason, 1988). Thus, for small scale land cover variability the relevant scales are the blending height L_b and u_* . A similar analysis as above yields as relevant horizontal length scales, L_x , 50 to 5.000 m from $L_b = 5$ to 500 m:

$$\frac{u_*}{u} \approx \frac{L_b}{L_x} \quad (2)$$

The concept of the blending height was formalized by Mason (1988) and assumes that at some height in the atmospheric boundary layer a quasi equilibrium exists between the balance of horizontal velocity advection and vertical flux perturbation:

$$u_0 \frac{\delta u}{\delta x} \approx \frac{\delta \Delta \tau}{\delta z} \quad (3)$$

where u is the windspeed and u_0 the undisturbed flow above the blending height and τ the Reynold's stress. This assumption implies that at this height

the fluxes are still approximately in local equilibrium whereas the windspeed represents almost an area average and has "blended out". Several refinements were suggested to this definition (e.g. Claussen, 1990, Wood and Mason, 1991), but its essence has remained the same. Straightforward substitution of length scales under the assumption that the velocity perturbation follows a vertically logarithmic distribution leads to an estimate of the blending height (Mason, 1988, Wood and Mason, 1991):

$$l_b = L_x \left(\frac{u_*}{u} \right)^2 \quad (4)$$

where L_x is a characteristic length scale for the horizontal variations. Several other estimates of the blending height may be used for practical purposes, such as the diffusion height scale or $L_x/200$ (Mason, 1988) or the height of the lowest model layer, but equation 4 provides arguably the best physical interpretation of the concept and will be used in the calculations presented in this paper. The concept of blending height has proven remarkably successful in a number of applications where its use in estimating effective parameters was compared against complex numeric simulations and a blending height for scalar transport was defined (Mason, 1988, Claussen, 1990, Wood and Mason, 1991 and Blyth *et al.*, 1993).

It is useful at this stage to define "effective parameters". Effective parameters (e.g. Fiedler and Panofsky, 1976) are parameters which give rise to the same flux as if that flux was calculated from individual patch contributions, each with their own parameter. Expressed more formally, using resistance notation this becomes for momentum:

$$\langle \tau \rangle = \rho \frac{u_a}{r_a} \quad (5)$$

and for heat and water vapour

$$\langle H \rangle = \frac{(\langle T_s \rangle - T_a) \rho c_p}{r_{ah}} \quad (6)$$

$$\langle \lambda E \rangle = \frac{(\langle e_s \rangle - e_a) \rho c_p}{(r_s + r_{aw}) \gamma} \quad (7)$$

where $\langle \rangle$ denotes an area average, τ the momentum flux, H the sensible heat flux, λE the latent heat flux, r a resistance with subscript ah and ae for heat and water vapour respectively, ρ the density of air, c_p the specific heat of air, λ the latent heat of vaporization, γ the psychrometer constant and T temperature, e water vapour pressure where the subscript s denotes a surface value and a a value at a reference level (blending height) in the air.

Locally, at point x , the following relationships still apply:

$$\tau(x) = \rho \frac{u_a}{r_a(x)} \quad (8)$$

$$H(x) = \frac{(T_s(x) - T_a) \rho c_p}{r_{ah}(x)} \quad (9)$$

$$\lambda E(x) = \frac{(e_s(x) - e_a) \rho c_p}{(r_s(x) + r_{aw}(x)) \gamma} \quad (10)$$

Area-averaging equations 8 to 10 and equating with equations 5 to 8 yields the following expressions for the effective resistances:

$$r_a^e = \frac{\langle \tau(x) r_a(x) \rangle}{\langle \tau \rangle} \quad (11)$$

$$r_{ah}^e = \frac{\langle H(x) r_{ah}(x) \rangle}{\langle H \rangle} \quad (12)$$

$$(r_{as} + r_p)^e = \frac{\langle \lambda E(x) (r_{as}(x) + r_p(x)) \rangle}{\langle \lambda E \rangle} \quad (13)$$

which show that the effective resistance in heterogeneous terrain is an *area-average weighted by flux* (e.g. Blyth *et al.*, 1993). This makes averaging resistances a non trivial task as a priori knowledge of the fluxes is required. It is also of interest to note that the aerodynamic and surface resistance for latent heat appear as a single term in the equations. It is possible to extend this analysis to the case of flux equations with roughness lengths (Wood and Mason, 1991), but the conclusions remain essentially the same.

NUMERICAL MODELS

Numerical models, based on the 2-D Bousinesq approximations of Navier-Stokes equations, have been used to study the effect of changes in surface cover (Mason, 1988, Claussen, 1991, Wood and Mason, 1991, Blyth *et al.*, 1993). A typical example of such a study (Claussen, 1990) is shown in Figure 2a, 2b and 2c. The roughness lengths for momentum follows a step change

from 10^{-3} m in the homogeneous domain to 10^{-2} m at the area $10\text{m} \leq x \leq 0\text{m}$; $\ln(z_{0H}/z_{0m})$ is set at 2.3. In figure 2a a deceleration of the flow can be observed above the strip as well as upstream and downstream of the change. Above the strip also an acceleration of the mean flow can be noticed. Both acceleration and deceleration are caused by perturbations of the mean pressure field (Claussen, 1990). For scalar concentration a similar picture emerges as shown in Figure 2b, albeit that concentration reductions are less severe (scalar transport is not directly affected by pressure) and extend to a smaller height than the flow reduction in Figure 2a. Wood and Mason (1991) found similar patterns resulting from their linear analysis of this flow problem. No enhancement of concentration is however observed. Figure 2c shows the combined effect of a change in roughness length and surface resistance for scalar flow. At $-10\text{m} \leq x \leq 0\text{m}$ a change in surface resistance is specified in the model of 10^5 m s^{-1} with the rest of the domain keeping a zero resistance. The picture now drastically changes with an enhancement of scalar concentration downstream of the step change. These results show the differences in the effect of changes in surface characteristics for scalar fluxes and concentrations and momentum, the latter changes being affected by pressure perturbations as well. Also they show that scalar perturbations do not reach as far in the boundary layer as do momentum perturbations. A problem in parameterization emerges in that the combined effect of a change in surface resistance and roughness is drastically different from the two individual changes. Similar conclusions were obtained by Wood and Mason (1991) and Blyth *et al.* (1993) and Dolman (1993b).

These numerical studies have been useful to look in detail at some of the processes relevant for aggregation. By dividing up the flow in a model resolved, grid scale part, and an unresolved part, the subgrid variability, Claussen (1991) was able to show the importance of several of the terms involving sub-grid correlations. In contrast to Mahrt's (1988) results he found that in the case of momentum and mass the terms containing subgrid correlations are small, compared to those of heat transfer. To a considerable degree the individual terms tend to cancel each other in the flow configuration Claussen used, but a non-negligible overall effect remains to be taken into account.

The main conclusions from the numerical models studies are that the velocity perturbations are generally of greater magnitude than the scalar perturbations and reach higher levels in the boundary layer, and that at a certain height the flow or concentrations become homogeneous. The latter conclusion gives credibility to the use of the blending height concept in averaging surface parameters such as roughness length and resistance.

Although these theoretical studies have provided great insight into the phenomena associated with aggregation, it is not immediately obvious whether similar conclusions will be reached for real life conditions. Kalthoff *et al.* (1993) have applied a three dimensional non-hydrostatic model of the atmosphere with an advanced, calibrated land surface model, to an area in central Germany, the Hildesheimer Börde to test certain averaging schemes. They also found perturbations of potential temperature and humidity to even out above a

certain height, the blending height. They then used various resolutions of the model to test whether simple averaging rules (Dolman, 1993a) could provide similar area average fluxes of heat and energy. Their main result is summarized in Figure 3, where a comparison of energy balance components between three models runs with resolution, 0, (the 1-D case), 450 and 990 m. The 1-D run applies standard aggregation rules (e.g. Dolman, 1993a) to calculate effective parameters for the full domain. The differences between the runs are surprisingly small, less than 10% for H , λE and the soil heat flux G . This result suggests that by applying realistic averaging rules (see below) it is possible to use simplified representations, effective parameters, in mesoscale models as long as the original surface information is sufficient. This conclusion is further based on the fact that the advective processes operating in such a heterogeneous area cancel each other to a sufficient extent to have a negligible effect on the area average fluxes.

PARAMETERIZATIONS

For the case of momentum roughness length Mason (1988) proposed to use the blending height (eq. 3 and 4) to estimate effective roughness length for an area of two surfaces (fractional coverage f_1 and f_2) with resistances z_{01} and z_{02} :

$$\frac{1}{z_0} = \frac{f_1}{\ln\left(\frac{l_b}{z_0}\right)^2} + \frac{f_2}{\ln\left(\frac{l_b}{z_{02}}\right)^2} \quad (14)$$

He showed that the momentum exchange was disproportionately affected by the large roughness elements in the domain. This result is explained if roughness length is converted to aerodynamic resistance (Blyth *et al*, 1993):

$$r_{am} = \frac{1}{k^2 U} \left[\ln\left(\frac{l_b}{z_0}\right) - \Psi_m \right]^2 \quad (15)$$

where k is van Karman's constant and Ψ_m the integral stability correction for momentum. At the blending height, the horizontal velocity is, by definition, constant across the domain. At the surface the no-slip condition implies that the velocity is always zero and consequently the effective resistance for momentum transport can be obtained by taking the parallel average of the resistances:

$$\frac{1}{r_a'} = \left\langle \frac{1}{r_a} \right\rangle \quad (16)$$

The effective resistance will thus be dominated by the smaller resistances in the domain, which correspond to high values of roughness length.

In contrast to momentum transport, transport of heat lacks a no slip condition at the surface and the calculation of an effective resistance to heat transport is complicated by the variability in surface temperature. Wood and Mason (1991) investigated this problem by imposing a constant heat flux over the domain. In this case the effective resistance,

$$r_{ah} = \frac{1}{ku_s} \left[\ln\left(\frac{z_o}{z_0}\right) - \Psi_h \right] \quad (17)$$

where Ψ_m is the integral stability correction for heat, is given by the average of the series resistances i.e.:

$$r_{ah}^e = \langle r_{ah} \rangle \quad (18)$$

Averaging procedures for heat and momentum are therefore not necessarily the same (Blyth *et al.*, 1993, Dolman, 1993b). Indeed as Wood and Mason (1991) show the behaviour of the effective roughness length for heat is dominated by the (thermally) smooth elements with low values of roughness length.

However, the use of the roughness length for heat is problematical in heterogeneous terrain. The roughness length for heat depends on the value of the gradient in temperature and the sensible heat flux and is thus a complex function of the surface energy balance in which aerodynamic properties interact with physiological surface or canopy resistances. This interaction is more appropriately expressed by equations of the Penman-Monteith type

(Monteith, 1981) which consist of a combination of the surface energy balance and the transfer equations. For latent heat:

$$\lambda E = \frac{\Delta A + \frac{p c \delta e}{r_{ah}}}{\Delta + \gamma \frac{(r_{ae} + r_g)}{r_{ah}}} \quad (19)$$

where separate resistances for heat and water vapour have been used and Δ is the slope of the saturated vapour pressure temperature curve.

Blyth *et al.* (1993) used the Penman-Monteith equation as the bottom boundary condition in the 2-D boundary layer model used by Mason (1988) and Wood and Mason (1991) to compare simple aggregation rules. They tested the application of the blending height (eq. 4) and simpler rules for averaging aerodynamic and surface resistance. The procedure for calculating effective parameters with the blending height was highly iterative to be able to take stability corrections into account and is outlined in detail in their paper (see also Wood and Mason, 1991). They also tested the application of the parallel and series average resistances (eq. 16 and 17) and developed a practical solution for "real life" aggregation which consisted of taking the average of the series and parallel averages:

$$r' = \frac{1}{2} \left[\langle r \rangle + \left\langle \frac{1}{r} \right\rangle^{-1} \right] \quad (20)$$

Table 1 shows the results of this comparison for surface resistance whilst Table 2 shows results for aerodynamic resistances to sensible and latent heat respectively for the blending height averaging only. If the range in surface

resistance is small or if the surface resistances are very high, the effective resistance is relatively insensitive to the averaging method. The values obtained by the blending height model, which uses values of temperature, windspeed and humidity at the blending height consistently are in better agreement with the 2-d model results than the other averaging methods, although the use of equation 20 qualifies as a good second option. Although these results are of a theoretical nature, they put limits on the aggregation problem in the sense that it is now possible to mark those vegetation types for which aggregation is possible or impossible.

The parametrizations or models described in this paper all assume statistical homogeneity within the domain, i.e. the variations are spread randomly across the domain. If the heterogeneity becomes less random directional effects may become important. None of the studies described here deals with this phenomenon.

THE ROUGHNESS LENGTHS FOR HEAT AND MOMENTUM

Wood and Mason (1991) were among the first to suggest that roughness lengths for heat and momentum needed to be aggregated in a different manner (see also Claussen, 1990, 1991). Figure 4 shows a comparison of effective roughness lengths derived from a numerical 2-D boundary layer model with those obtained by applying their theory of blending height. This graph shows the different behaviour of the effective roughness lengths for heat and momentum in heterogeneous terrain for a range of stabilities. It also shows the ability of the blending height model to correctly estimate the effective values. For neutral conditions the values of z_{om} are enhanced, typically by a factor 2, whereas the values of z_{oh} are reduced by a factor 3. The value denoted as average z_{oa} is obtained from $\log(z_{oa}) = \langle \log(z_{zo}(x)) \rangle$. Wood and Mason (1991) concluded from this figure that for most practical applications the effects of stability on the value for effective momentum roughness length was unlikely to be of significance, whereas for z_{oh} this effect could have some significance.

In heterogeneous terrain the sources of latent, sensible heat and momentum are not necessarily the same. Blyth and Dolman (1994) show how in the case of sparse crops (with length scales $L_x < 100\text{m}$) this effect can be dealt with elegantly by parameterizing the laminar boundary layer resistances, which are felt only by scalar transport, explicitly (Owen and Thompson, 1963).

Following the analysis presented in figure 1, and the success of the blending height model in estimating effective parameters, it is now possible to estimate the effect of spatial heterogeneity on a larger scale to predict effective values for the resistances and roughness lengths for heat and momentum. It is important to note that the point of interaction of the two surfaces (at $z_0 + d$ in the sparse crop case) is in this case moved to a level above the canopy, at the blending height, which is calculated from equation 4. However, the basic framework for solving the energy balance equations of the two surfaces remains similar to that of the sparse crop model and is shown in figure 5 (Dolman, 1993b). The model thus consists of two components: one to solve the combined energy balance equations of two surfaces (Shuttleworth and Wallace, 1988 and Dolman, 1993b) and the second consists of the application of the blending height theory to calculate the level above the canopy at which the interaction between the surfaces takes place.

Unlike the sparse crop model, where the difference in momentum and heat transport is parameterized by laminar boundary layer resistances (r_b^u, r_b^c) it is assumed in the case of the present larger scale heterogeneity that the difference can be parameterized by simply taking $\ln(z_{om}/z_{or})=2.3$, as for homogeneous vegetation and calculate a resistance from the canopy level to the blending height. The aerodynamic resistance r_a^a is then calculated from the blending height to the reference or first model layer. Effectively it is assumed that both patches are sufficiently homogeneous to justify this assumption (see Garrat, 1992 for an analysis of $\ln(z_{om}/z_{or})$ for homogeneous vegetation).

Figure 6 shows the effect of changes in the partial cover in such a large scale sparse crop model ($L_x = 250\text{m}$). In the case of no variation in surface resistance, r_s , and small variation in momentum roughness length for the individual patches, the variation of $\ln(z_{om}/z_{o1})$ with fractional cover is small. Changing r_s for each of the patches, or making the difference between the roughness length larger, changes the energy partitioning of the patches and the average fluxes of sensible and latent heat now bear a different relation to the average gradients. The ratio of the two roughness lengths is now much more variable with the fractional surface cover.

Figure 7 shows how radically different the relation of the effective to the average roughness length is for temperature and momentum. This figure shows good agreement with the results obtained by Wood and Mason (1991) under slightly different assumptions. For momentum there is a rapid initial increase in effective roughness length, caused by the introduction of a small number of high roughness elements in the domain. By contrast, when low roughness elements are introduced in a high roughness length environment, the effective roughness length does not change dramatically. For the temperature roughness length an opposite pattern is obtained. Here a decrease in effective roughness length is observed when thermally rough elements are introduced in an domain consisting of thermally smooth elements. Blyth and Dolman (1994) concluded that for sparse canopies it is better to use a surface energy balance model which explicitly models the different sources of heat momentum and water vapour. A similar conclusion may be drawn for areas with large heterogeneity.

DISCUSSION

One of the problems in applying the aggregation rules suggested in this paper concerns the length scales of variation. Blyth (1994) shows how the sparse canopy model used above can be employed to estimate effective parameters at a range of spatial scales. The scheme employed was similar to the scheme outlined in Figure 1. Figure 8 shows how the ratio of the two resistances changes with length scale of the variation. The solid line predicts this variation from the blending height theory whilst the circles represent simulations with a 2-D numerical boundary layer model. The agreement is surprisingly good, suggesting that the concept of blending height, coupled with adequate descriptions of the land surface, can reliably estimate surface fluxes from heterogeneous surface cover at a range of length scales.

In principle by using the (sparse crop) surface energy balance models at larger scales than for which they were originally developed-as shown in the previous section of this paper- the need for deriving effective values vanishes altogether. If the separate fluxes can be efficiently calculated from a such a model, by taking into account the basic interaction process (advection), this method should be used in (meso scale) meteorological modelling rather than a method which calculates effective parameters in advance by applying aggregation rules, as the latter will always involve some level of compromise and a consequent loss of accuracy. It is important to note that the use of a tile or mosaic model (Koster and Suarez, 1992) also neglects the interaction between

the patches if the lowest model layer is not sufficiently close to the surface (i.e. is higher than the blending height) and also does not always produce flux rates similar to those calculated by the model presented in the previous paragraph.

It would be relatively straightforward to include simple parametrizations of edge effects or changes in zero plane displacement (Claussen and Klaassen, 1993) in this formulation, but the practical significance of these effects still needs to be established when working at large scales. This can be seen also in Figure 8 where the point most left on the graph represents a *measured* value related to a vegetation type in the Sahel, tiger bush, which consists of areas of bare soil interspersed with vegetation (see Blyth, 1994) and consequently would suffer from large edge effects. The proximity of this point to the theoretical line, calculated without parameterizing edge effects, suggest that these effects may be small, at least for this type of edge effect.

This raises the issue of validating these theoretical approaches. Although some validation work was presented in this paper, the approaches suggested require more thorough validation on certain specific points. The methodology developed in the large scale field experiments (Shuttleworth, 1988) from IGBP and WCRP would then probably be in need of some refocussing to address these issues and to align the experiments in more detail with modelling efforts.

RECOMMENDATIONS

Simple averaging of the series and parallel resistances is a good option to estimate effective parameters for resistances. When more detail is required the use of a blending height in averaging is preferred. Resistances for heat and water vapour need to be averaged in the context of an surface energy balance constraint as they are strongly dependent on the interplay between surface parameters and overlying meteorology.

A multiple source energy balance model can easily be adapted to take simple heterogeneity into account without adding extra computational requirements.

QUESTIONS

Question 1, dr. A. Jochum, DLR, Germany.

Do you have any specific ideas on how to validate the suggested aggregation rules leading to a modified design of large scale field experiments in the future?

Answer

In the past the design of field experiments like HAPEX-Mobilhy, EFEDA and HAPEX-Sahel has been more on the basis of availability of field measurements and less on the data requirement of modellers. The theories presented in this paper can in principle be tested and one way of doing that would be to run models *before* an experiment on the basis of existing data and test the basic sensitivities and develop from there an observational strategy to validate essential parts and predictions of the models. The upcoming experiment in the Amazon basin should provide ample opportunities to do so.

Question 2 Dr. M.R. Raupach, CSIRO, Australia.

Presumably there is spectrum of blending heights corresponding with the spectrum of length scale variation in surface properties. How would we describe this spectrum?

Answer

Such a spectrum would in all likelihood exist. Spectral analysis of remote sensing images may be the way forward to determine the shape of these spectra. At this stage the most probable solution to the problem is to classify the length scales along the lines suggested in this paper. A natural upper limit exists at, say, 10 km, as from there on the boundary layer exhibits a different growth rate and the basic assumptions leading to the blending height are not fulfilled any more. Using only one or two blending heights simplifies the problem dramatically and is probably justified by the fact that it is an "order of magnitude" estimate. Some amount of error in the exact height is therefore allowed and does not lead always to noticeable differences in the calculated flux rates.

$R_1 R_2 \text{ (sm}^{-1}\text{)}$	Model	L_0 -model	Eq. 16	Eq. 18	Eq. 20
$Z_{01} Z_{02} \text{ (m)}$					
60/120	84	84	80	90	80
0.01/1					
0/120	30	28	0	60	30
1/1					
0/60	25	23	0	30	15
0.01/1					
300/600	383	381	400	450	425
0.01/1					
600/300	453	437	400	450	425
0.01/1					
0/120	39	36	0	60	30
0.01/1					
120/0	36	27	0	60	30
0.01/1					

Table 1 Effective surface resistance (sm^{-1}) for a number of cases where the surface resistance and momentum roughness length vary within a single domain calculated with a numerical model, the blending height (L_0) model and three averaging equations.

$R_1 R_2$	Latent heat model	Sensible heat model	Latent heat L_0 -model	Sensible heat L_0 - model
$z_{01} z_{02}$				
60/120	38	30	41	30
0.01/1				
0/120	19	10	23	9
1/1				
0/60	45	38	42	41
0.01/1				
300/600	42	30	44	30
0.01/1				
600/300	38	33	31	33
0.01/1				
0/120	44	24	49	23
0.01/1				
120/0	29	155	24	238
0.01/1				

Table 2 **Effective aerodynamic resistance (sm^{-1}) for a number of cases**
where the surface resistance and momentum roughness length
vary within a single domain calculated with the blending height
model and numerical model.

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LIST OF CAPTIONS.

Figure 1. Schematic picture of a sparse canopy model (a) and a version used for larger scale heterogeneity (b).

Figure 2a. Isolines of percentage relative difference between perturbed local and unperturbed upstream horizontal mean velocity. Flow configuration $z_0 = 10^{-2}$ m for $-10 < x < 0$, elsewhere $z_0 = 10^{-3}$ m (redrawn after Claussen, 1990).

Figure 2b. Isolines of percentage relative difference between perturbed local and unperturbed upstream scalar concentration. Flow configuration $z_0 = 10^{-2}$ m for $-10 < x < 0$, elsewhere $z_0 = 10^{-3}$ m, $\ln(z_{0m}/z_0) = 2.3$, $r_s = 0 \text{ sm}^{-1}$ (redrawn after Claussen, 1990).

Figure 2c. Isolines of percentage relative difference between perturbed local and unperturbed upstream scalar concentration. Flow configuration $z_0 = 10^{-2}$ m for $-10 < x < 0$, elsewhere $z_0 = 10^{-3}$ m and $r_s = 10^5 \text{ sm}^{-1}$ at $-10 < x < 0$ (redrawn after Claussen, 1990).

Figure 3. Comparison of energy balance components of three model runs with different resolution. Thick solid line $\lambda E_{450} - \lambda E_{990}$, long short long dashed $H_0 - H_{990}$, thin solid line $G_0 - G_{990}$, thick dashed line $G_{450} - G_{990}$, dotted line $\lambda E_0 - \lambda E_{990}$, thin dashed line $H_{450} - H_{990}$.

where λE is evaporation, G soil heat flux, H sensible heat flux and the subscripts refer to the resolution of the land surface maps. (redrawn after Kalthoff et al., 1993)

Figure 4. Comparison of effective roughness length for heat and momentum calculated with the 2-D numerical model (crosses) using Businger-Dyer stability corrections, triangles (KEYPS profiles) and the blending height model (lines) (redrawn after Wood and Mason, 1991).

Figure 5. Resistance diagram for a sparse crop surface energy balance model (redrawn after Dolman, 1993b).

Figure 6. Variation of $\ln(z_{0m}/z_{0t})$ with fractional cover for three resistance configurations with fractional surface cover (solid line $r_s^1 = r_s^2 = 50 \text{ s m}^{-1}$, $z_{01} = 1 \text{ m}$, $z_{02} = 0.01 \text{ m}$, long dashed line $r_s^1 = 50 \text{ s m}^{-1}$, $r_s^2 = 500 \text{ s m}^{-1}$, $z_{01} = 1 \text{ m}$, $z_{02} = 0.01 \text{ m}$, short dashed line $r_s^1 = 50 \text{ s m}^{-1}$, $r_s^2 = 50 \text{ s m}^{-1}$, $z_{01} = 0.1 \text{ m}$, $z_{02} = 0.01 \text{ m}$).

Figure 7. Variation of the ratio of effective to average roughness length for heat and momentum with fractional surface cover for $r_s^1 = r_s^2 = 50 \text{ s m}^{-1}$, $z_{01} = 1 \text{ m}$, $z_{02} = 0.01 \text{ m}$ (solid line temperature, long dashed momentum) and $r_s^1 = 50 \text{ s m}^{-1}$, $r_s^2 = 500 \text{ s m}^{-1}$, $z_{01} = 1 \text{ m}$, $z_{02} = 0.01 \text{ m}$ (medium dashed line temperature, short dashed momentum).

Figure 8. Ratio of resistance from blending height to model layer to that of canopy level to blending height (see Figure 5 and text for explanation of resistance network). The solid line shows values of the ratio calculated by the blending height theory, the triangle is a "measured value" for Sahelian tiger bush, the closed circles represent 2-D numerical simulations.

15-17

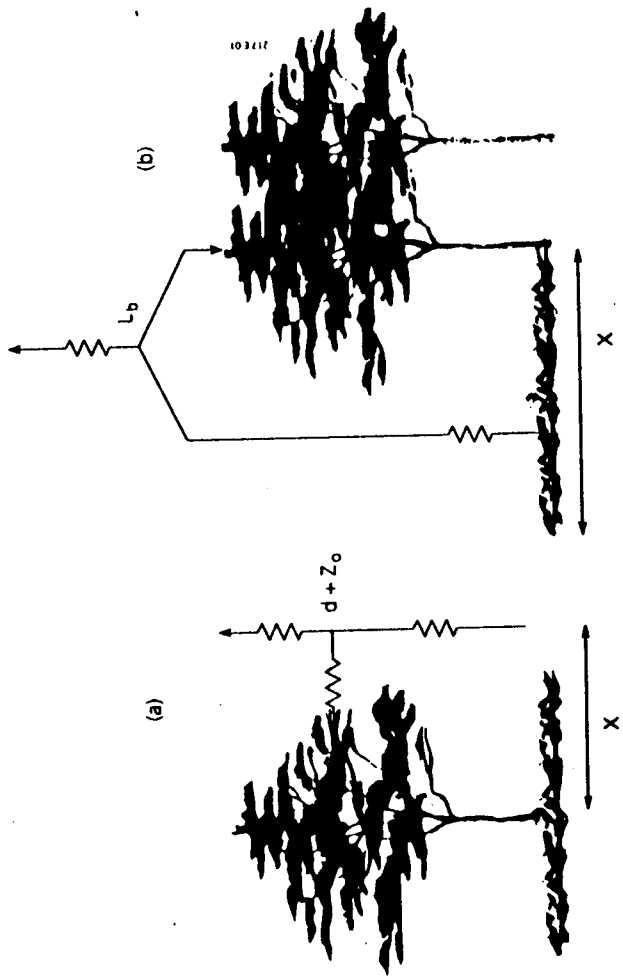


Fig 2

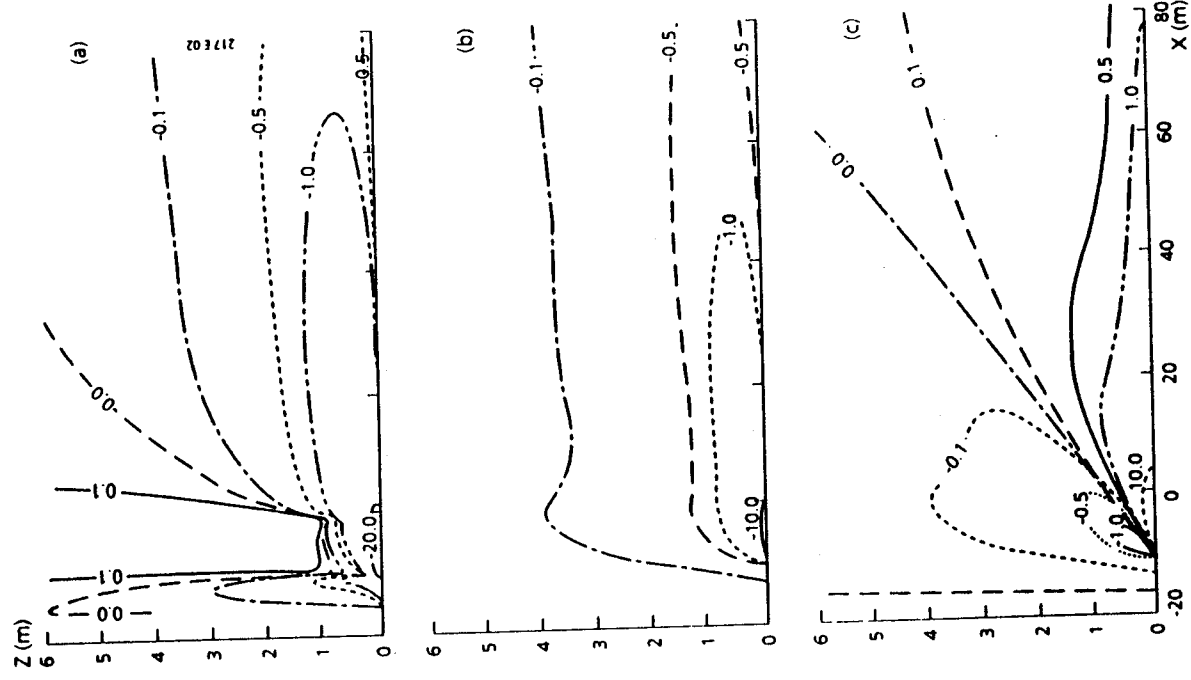


Fig 3

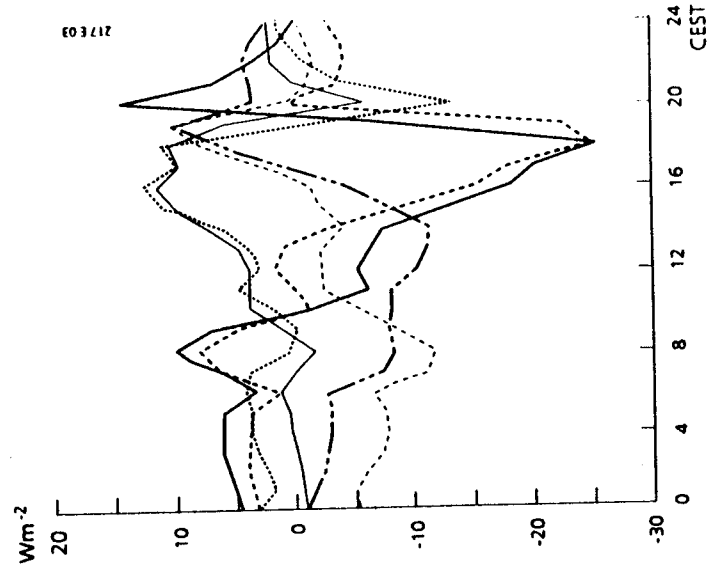


Fig 5

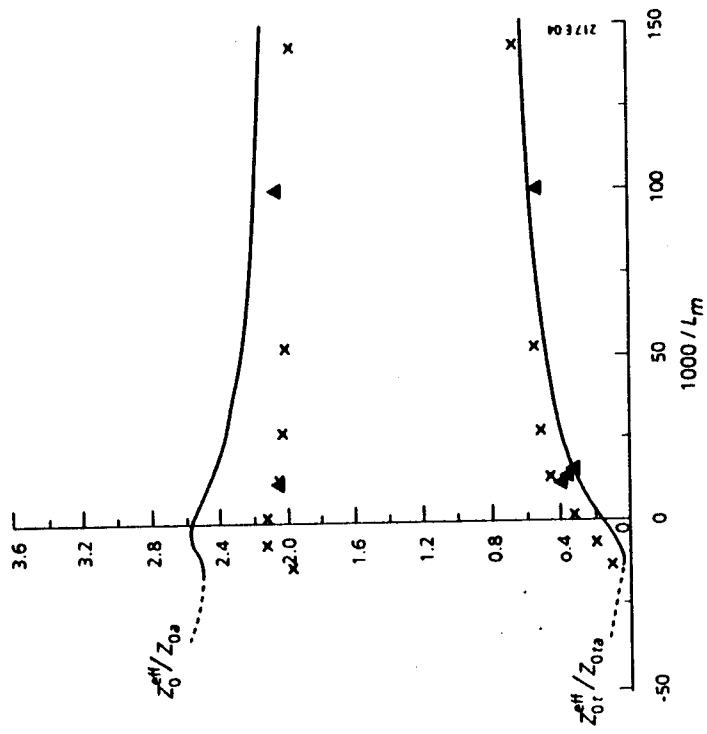


Fig 5

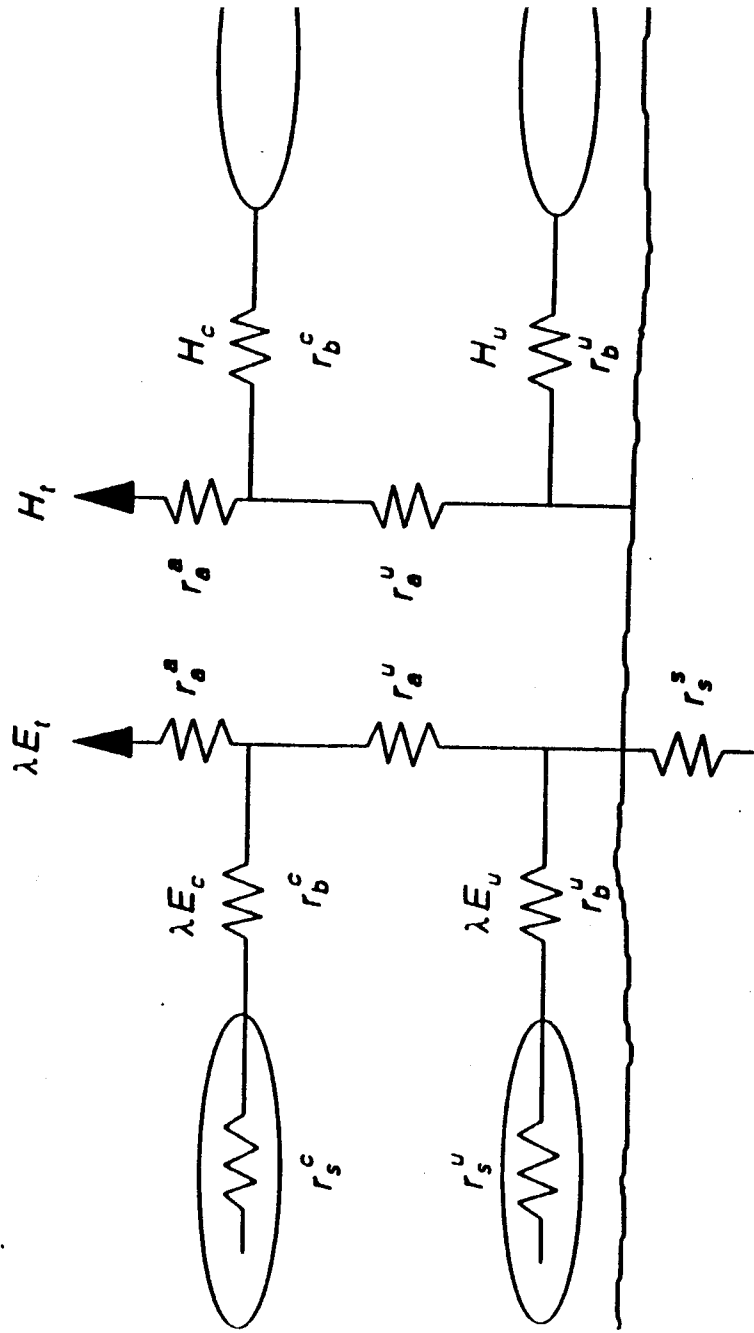
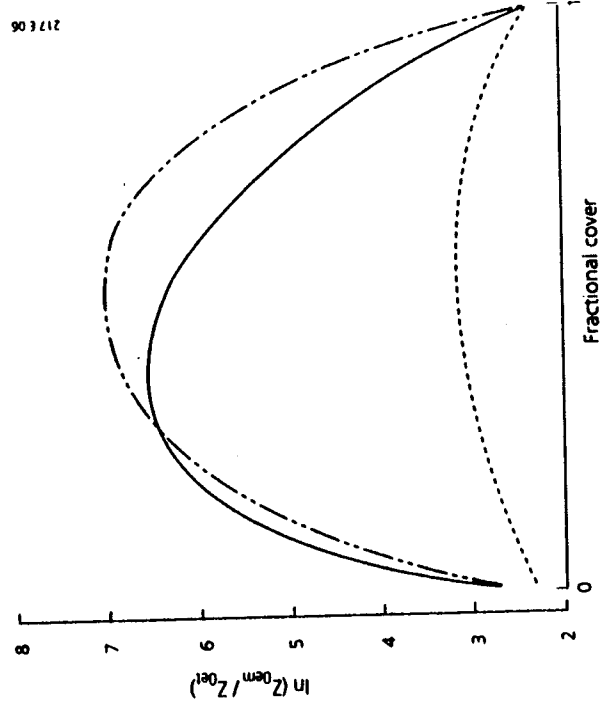


Fig 6



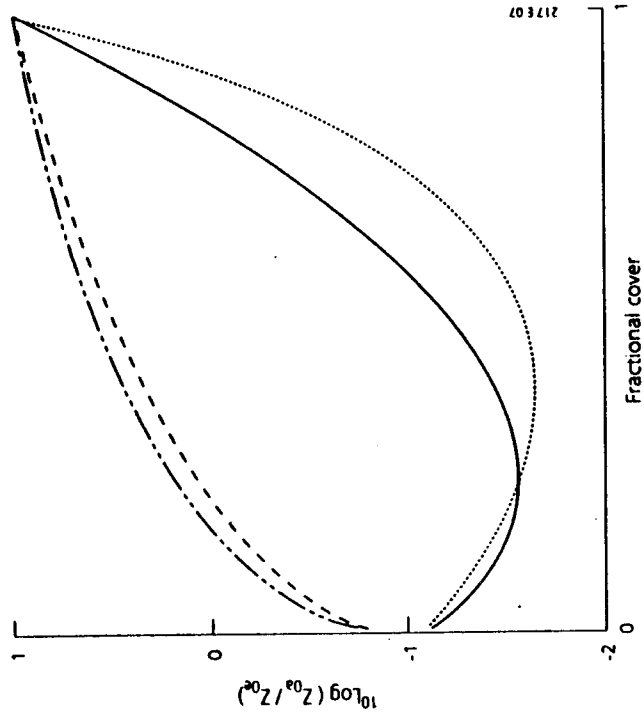
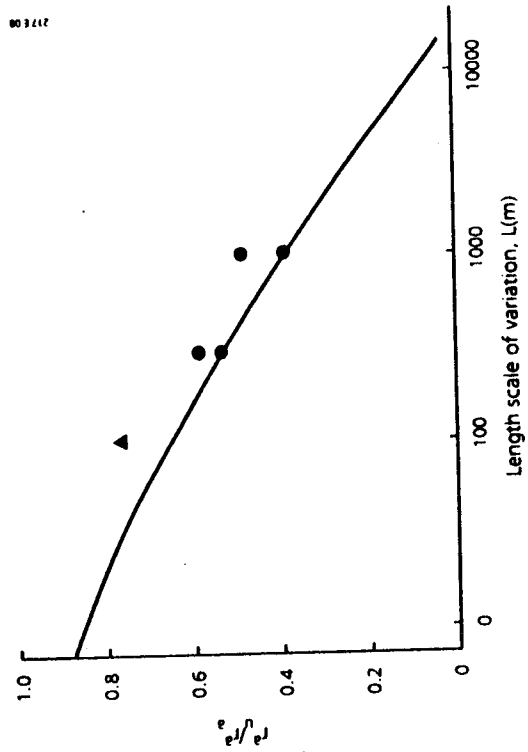


Fig 7

Fig 8



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The scaling characteristics of soil parameters:
From plot scale heterogeneity to subgrid parameterization

Paper presented at the IGBP-BAHC, GEWEX-ISLSCP Workshop on Aggregation,
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Abstract

The variation in soil texture, surface moisture or vertical soil moisture gradient in larger scale atmospheric models may lead to significant variations in simulated surface fluxes of water and heat. The parameterization of soil moisture fluxes at spatial scales compatible with the grid size of distributed hydrological models and mesoscale atmospheric models ($\sim 100 \text{ km}^2$) faces principal problems which relate to the underlying microscopic or field scale heterogeneity in soil characteristics.

The most widely used parameterization in soil hydrology, the Darcy-Richards (DR) equation, is gaining increasing importance in mesoscale and climate modelling. This is mainly due to the need to introduce plant-interactive soil water depletion and stomatal conductance parameterizations and to improve the calculation of deep percolation and runoff. Covering a grid of several hundreds of square kilometres, the DR parameterization in such soil-vegetation-atmosphere-transfer schemes (SVATs) is assumed to be scale-invariant. The parameters describing the non-linear, area-average soil hydraulic functions in this scale-invariant DR-equation should be treated as *calibration-parameters*, which do not necessarily have a physical meaning. The saturated hydraulic conductivity is one of the soil parameters to which the models show very high sensitivity. It is shown that saturated hydraulic conductivity can be scaled in both vertical and horizontal directions for large flow domains.

In this paper, a distinction is made between *effective* and *aggregated* soil parameters. Effective parameters are defined as area-average values or distributions over a domain with a single, distinct textural soil type. They can be obtained by scaling or inverse modelling. Aggregated soil parameters represent grid-domains with several textural soil types. In soil science dimensional methods have been developed to scale up soil hydraulic characteristics. With some specific assumptions, these techniques can be extrapolated from classical field-scale problems in soil heterogeneity to larger domains, compatible with the grid-size of large scale models. Particularly promising is the estimation of effective soil hydraulic parameters from area averaging measurements through inverse modelling of the unsaturated flow.

Techniques to scale and aggregate the soil characteristics presented in this paper qualify for direct or indirect use in large scale meteorological models. One of the interesting results is the *effective* behaviour of the *reference curve*, which can be obtained from similar media scaling. If the conclusions of this paper survive further studies, a relatively simple method will become available to parameterize soil variability at large scales. The inverse technique is found to provide *effective soil parameters* which perform well in predicting *both* the area-average evaporation and the area-average soil moisture fluxes, such as subsurface runoff. This is not the case for *aggregated soil parameters*. Obtained from regression relationships between soil textural composition and hydraulic characteristics, these *aggregated* parameters predict evaporation fluxes well, but fail to predict water balance terms such as percolation and runoff. This is a serious drawback which could eventually hamper the improvement of the representation of the hydrological cycle in mesoscale atmospheric models and in GCMs.

1. Introduction

Surface fluxes simulated by mesoscale atmospheric models are sensitive to changes in soil texture, surface moisture or vertical soil moisture gradient (e.g. Mahtouf et al., 1987, McCumber and Pielke, 1981; Noilhan and Lacarrere, 1995). Johnson et al (1991) showed that inclusion of sub-grid spatial variability of soil moisture and precipitation in a General Circulation Model (GCM) results in a significant improvement of the modelled hydrological balance. These and other studies (Rowntree, 1991) indicate a need to parameterize sub-grid variability in soil parameters and soil moisture behaviour in mesoscale and climate models.

Parameterization of soil moisture fluxes at these scales ($\sim 100 \text{ km}^2$) faces principal problems which relate to the heterogeneity of soil characteristics that exists at microscopic, plot and field scales ($10^{-2} - 10^3 \text{ m}$; e.g. Beven, 1994; Russo and Bresler, 1980). For instance, the widely used Darcy-Richards (DR) approach assumes that the pressure at a particular point in the flow domain is in equilibrium over some "representative elementary volume" (REV) and that the water movement takes place in response to a pressure gradient at REV scale. At the local scale, this assumption is generally accepted. However, at the larger scale this concept is often questionable due to the strong non-linear relationship between the moisture fluxes and soil water pressure gradients (Beven, 1994). Also, relationships between the effective value of a parameter and the underlying within-grid heterogeneity may vary with different flow conditions. There is an intrinsic relationship between the scale and the underlying heterogeneity of the soil characteristics. Some studies assume the effective parameter concept to be valid for grid scales that are larger than the scale of individual catchments (Refsgaard et al., 1992). However, such approaches cannot be justified by any theory for the estimation of effective parameters and therefore remain a convenient hypothesis (Beven, 1994).

The key hydrological flow processes active at these small scales are generally neglected or unresolved at the grid scale of distributed hydrological models or mesoscale atmospheric models. Operational mesoscale atmospheric models have a typical grid size between 20 - 50 km. This means that a grid point represents an area that usually contains several pedological soil types (e.g. sandy and clay soils), or an area that contains more than one "functional ensemble" of textural classes (light, medium and heavy soils). The scaling and aggregation of soil hydraulic characteristics for mixed soil types remains largely un-explored, despite the fact that the pedosphere is basically a continuum and its classification into discrete soil types is somewhat arbitrary (Scholes et al, 1995).

High resolution (mesoscale) models account for some soil heterogeneity through the variation of soil characteristics over the grid elements of the model. This merely shifts the problem to estimation of the area-average values of soil hydraulic characteristics at a smaller grid-element scale (Noilhan and Lacarrere, 1995; Wood, this issue). In the study by Noilhan and Lacarrere such aggregated soil parameters were derived for a large mesoscale domain, based on the dominant soil texture class within each grid box of $10 \times 10 \text{ km}$. An aggregated soil texture class for the entire simulation domain was then created through averaging of the fractional representation of sand and clay within this domain. Different representations of the "domain-average" soil texture (effective versus dominant) were found to have a large influence on predicted mesoscale evaporation fluxes.

Numerical experiments with distributed hydrological models suggest that grid-scale effective values can - under certain conditions - provide acceptable results for the prediction of water fluxes (Binley et al, 1989, Feddes and Kabat, 1993). Later in this paper, a case study (Feddes et al. 1993a, 1993b) will be presented that demonstrates the successful use of an inverse modelling approach to estimate effective soil hydraulic parameters of a small size catchment. Studies where the concept of area averaged soil hydraulic properties is tested in the framework of a coupled hydrological-atmospheric mesoscale model are still scarce.

The difficulty of field and grid-scale parameterizations of soil hydrological processes lies not only in the soil heterogeneity and non-linearity of the unsaturated flow equations, but also in a mismatch between the scale of measurements and the scale of model prediction. This is perhaps little surprising, knowing that other surface fields, like for example the heat fluxes or surface conductances, suffer from the same handicap. However, the representative scale to which a single measurement in the soils relates is much smaller than that for the atmosphere. The soil is a semi-static medium in which laminar flow and small mixing ratios prevail, whereas in the atmosphere turbulent flow results in effective mixing at relatively large scales. It is therefore difficult to obtain large scale-integrated measurements of soil moisture. The range of techniques used to monitor in-situ soil water content (Kabat and Beekma, 1994) give point measurements of soil moisture. These point measurements are usually unrepresentative of larger areas and serve only to highlight the underlying variability in soil hydrological processes (Kabat et al, 1995a).

This problem is most urgent in climate modelling, where the difference between the scale of soil moisture measurements and model output is the largest. Therefore a need exists for techniques that measure soil hydrological and water balance variables at large spatial scales. One such integrated measure of the catchment hydrological response is the discharge measured at the catchment-outlet. Other, potentially useful "scale-integrators" of near-surface soil moisture are the airborne and satellite passive and active microwave techniques (Schmugge et al, 1992; Bastiaanssen et al, 1994).

With some exceptions, the scaling and aggregation methods, reviewed in this paper, are not entirely new and some have been widely applied in the field of soil hydrology and soil science. However, it seems that within the meteorological and climate modelling community these techniques have not gained appropriate attention. This paper makes an attempt to evaluate the relative merits and applicability of these techniques in the context of atmospheric mesoscale and climate modelling.

Within a single review paper, it is impossible to treat all existing methods for scaling up heterogeneous soil characteristics. This paper focuses mainly on the Darcy - Richards approach and discusses different possibilities of implementing this approach into parameterizations of heterogeneous soil hydrological processes in large-scale atmospheric and hydrological models. Other simplified methods, such as pseudo-steady soil water flow and the capacity approach (e.g. Kabat and Feddes, 1995) or the matrix flux potential method (e.g. ten Berge et al, 1987) are not discussed. It is however recognized that these simplified methods may in some situations have advantages over the use of Darcy-Richards model, for example

under the conditions of soil-data scarcity.

In this paper a distinction is made between *effective* and *aggregated* soil parameters. The effective parameters are defined as area-average values or distributions over a domain with a single, distinct pedological soil type. Aggregated soil parameters will represent grid- domains with several soil types. These definitions are somehow arbitrary and they are introduced in this paper mainly to make a distinction between estimation techniques of these soil parameters. In contrast with aggregated parameters, the effective parameters are obtained by extrapolating parameterizations (relationships) rather than measurements.

Section 2 deals with some theoretical aspects of the scale-invariance of the Darcy-Richards equation, which is the most widely used parameterization in soil hydrology and which is gaining increasing importance in mesoscale and climate modelling. Section 3 elaborates on techniques to average saturated hydraulic conductivity which is one soil parameter to which models show a very high sensitivity. Dimensional methods to scale up soil hydraulic parameters from classical field-scale problems in soil heterogeneity to larger domains, are treated in section 4. Section 5 introduces large-scale inverse modelling as a means to obtain effective soil hydraulic parameters. Finally, in section 6 the aggregated soil parameters of the spatial grids which are composed of several soil types are analyzed through a numerical experiment. In the discussion the relative merits of these techniques are discussed.

2. Theory: is the Darcy - Richards equation scale invariant ?

Dealing with the problem of spatial heterogeneity in soils it is essential to revisit the most common parameterization framework of soil water flow, the Darcy-Richards equation :

$$\frac{1}{C(h_m)} \frac{\delta}{\delta z} \left[K(h_m) \left(\frac{\delta h_m}{\delta z} + 1 \right) \right] - \frac{S}{C(h_m)} = \frac{\delta h_m}{\delta t} \quad (1)$$

where h_m is soil matrix potential or pressure head, z is the vertical coordinate (positive upward), $K(h_m)$ is the hydraulic conductivity as a function of the matrix potential t is the time and $C(h_m)$ is differential soil water capacity, defined as the slope of the soil water retention curve $\delta\theta/\delta h_m$, with θ being the volumetric soil moisture content. S in Equation (1) is defined as a sink or source term to account for root water uptake or preferential fluxes in the soil matrix (Feddes et al, 1988). Equation (1) is a combination of a flux- potential gradient parameterization (Darcy equation) with conservation of mass (water) in the soil. Except for a limited number of special flow-situations (Raats, 1983), it can only be solved numerically by specifying initial and boundary conditions. The solution requires *a priori* parameterization of the soil hydraulic characteristics, i.e. the soil water retention and hydraulic conductivity functions.

As already pointed out by Beven (1994), Zijl and Lambert (1989), among many others, the hydrological modelling community continues to rely on the Darcy - Richards equation used with effective soil hydraulic functions, assuming that this is valid at all scales of application. The atmospheric modelling community also shows an increasing tendency to incorporate (quasi)-Darcy-Richards parameterizations in

soil-vegetation-atmosphere-transfer (SVAT) models. This is mainly due to the need to introduce plant-interactive soil water depletion and stomatal conductance parameterizations and to improve the calculation of deep percolation and runoff. Again, the Darcy-Richards parameterization is taken to be scale-invariant in such SVAT-schemes, which are developed for large scale applications such as in GCMs.

Can the scale-invariance of DR-equation be justified? To investigate the appropriateness of this assumption, the following theoretical framework is adopted (Whitaker, 1986; Zijl and Lambert, 1989, amongst others).

The main driving force in Equation (1), is the soil matrix potential gradient which can be expressed as :

$$\frac{\delta H}{\delta z} = \frac{\delta p}{\delta z} - \beta g \quad (2)$$

where $H = h_m + z$ presents the total soil water potential (hydraulic head), p is the total pressure of soil water, β is density of water and g is acceleration due to gravity. After multiplication by differential moisture capacity and neglecting the sink/source term, Equation (1) can now be written as:

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta z} K \left(\frac{\delta p}{\delta z} - \beta g \right) \quad (3)$$

This equation represents a simplified form of DR-parameterization. The air pressure in the unsaturated zone is equal to atmospheric air pressure. The total pressure of soil water in the unsaturated zone can now be calculated from :

$$p(x, t) = p_{atm}(t) - h(x, t) \quad (4)$$

with x the coordinate of a point in space, t the time coordinate and h being the capillary pressure (negative), known as soil water suction. Denoting area-average values by $\langle \rangle$, it is possible to write the equation for total soil water pressure on area-average basis:

$$\langle p \rangle(X, t) = p_{atm}(t) - \langle h \rangle(X, t) \quad (5)$$

where X is the coordinate at the centre of the area over which we are averaging. Replacing local soil water pressure in Equation (3) by the area-average pressure, we arrive at the following expression for the large-scale, area - average DR parameterization :

$$\frac{\delta \langle \theta \rangle(X, t)}{\delta t} = - \frac{\delta}{\delta z} \left[\langle K \rangle(X, t) \cdot \left\{ \frac{\delta \langle h \rangle(X, t)}{\delta z} + \langle \beta \rangle g \right\} \right] \quad (6)$$

with Z the centre of the soil profile depth. The area average values of the soil moisture content θ and of the hydraulic conductivity K are dependent on the local-scale solution of the soil suction $h(x, t)$:

$$\langle \theta \rangle(X, t) = \langle \theta(h(x, t)) \rangle \quad (7)$$

$$\langle K \rangle(X, t) = \langle K(h(x, t)) \rangle \quad (8)$$

However, because $\theta(h)$ and $K(h)$ are non-linear functions of h , the averages of the local values of θ and K , calculated from local pressures, do not equal regional values of θ and K , calculated from area averaged pressures:

$$\langle \theta(h(x,t)) \rangle \neq \theta(\langle h \rangle(X,t)) \quad (9)$$

$$\langle K(h)(X,t) \rangle \neq K(\langle h \rangle(X,t)) \quad (10)$$

It can be shown that even when $\theta(h)$ and $K(h)$ are approximated by linear functions, the variation of these functions from place to place within an area prevents a simple deterministic solution of a large-scale Darcy-Richards equation (Equation 6). Stochastic parameterization approaches have been therefore developed (Mantoglou and Gelhar, 1987a; Yeh et al, 1985) to overcome this problem. The overall complexity of these stochastic approaches however deters their direct implementation into parameterization schemes of the larger scale (coupled) atmospheric and hydrological models (see also Chen et al, 1994)

The non-stochastic averaging is therefore only possible, when the local small-scale solutions for $h(x,t)$ are known. The area - average, large scale soil water retention and hydraulic conductivity curves can be estimated from the local-scale ones, if a specific functional relationship between the area-average and the local solution for h is known:

$$h_{area}(x,y,z,t) = \int_{x,y} F(h_{loc}(x,y,z,t)) \quad (11)$$

In general, the function F is difficult to determine because it depends on the ratio between horizontal and vertical length of a flow domain. F also depends on the presence of horizontal flow and other factors such as the flow being stationary. The pressure variation in the horizontal plane can be neglected only in special cases, thus allowing F to be approximated by unity. An example of this are laboratory columns used to measure retention and conductivity curves (Kabat and Beekma, 1994), where the cross sectional area normal to the flow (x,y -plane) is relatively small compared to the length of the column in vertical direction. This means that the $h(\theta)$ and $K(h)$ relationships obtained through laboratory measurements are *effective values related to a certain volume of the soil column*. For the large flow domains of mesoscale hydrological and atmospheric models, the vertical dimension (i.e. depth of the soil profile) is several orders of magnitude smaller than the horizontal dimension and F is not known *a priori*. Equation (11) also shows that generally F will not be a unique function, because identical values of $\langle h \rangle$ can be obtained from many combinations of $h(x,t)$.

Summarizing, an area-average solution for DR-equation, which would retain the physical meaning of the soil hydraulic characteristics (i.e. soil water retention and hydraulic conductivity curves) remains computationally laborious, non-trivial and in many cases an almost impossible task, requiring an extended knowledge of local-scale information. Some specific solutions, based on a stochastic spatial averaging of the DR equation have been developed (Chen et al, 1994) to deal with some of these aspects.

Comparing aggregation principles as discussed above for soil water flow with

general averaging rules for meteorological variables as reviewed by Dolman and Blyth (this issue), some similarities are obvious. In both cases, the area average solution depends on *a priori* knowledge of a relationship between the local scale and area-average fluxes. However, there are also striking differences. Strong mixing in the atmosphere is a reason for almost continuous recovery of the characteristic large scale, synoptic profiles of air temperature, wind speed and humidity or water vapour pressure. It is then possible to define a reference height at which a single variable is well mixed (for example the blending height; see Dolman and Blyth, this issue) and to express a large scale (area average) single driving gradient with respect to this mixed variable. Mixing in soils is far less effective. Therefore a suitable mixing variable, which may be used in a way analogous to the aggregation of meteorological vertical fluxes, cannot be defined. Except for very special flow situations, such as steady state saturated vertical flow, none of the variables in Equation (6) can be "homogenized" at a theoretically pre-determined depth in the soil profile. In the expression for the large-scale, effective soil water pressure (Equation 5), the air pressure in the unsaturated zone is assumed to be scale invariant; however such an assumption does not hold for the capillary pressure. For special cases, like the capillary pressure at phreatic level (the level of the groundwater depth in the soil profile), the condition of scale-invariance of the total water pressure at a certain pre-determined depth would hold. But due to in-efficient vertical mixing even over a depth of a single soil layer (~ 1 m), a single gradient of the driving potential (pressure) does not have a clear physical meaning. Therefore the averaging needs to take into account the vertical distribution of the gradients and of the fluxes. Hence, in addition to the local scale solution in a horizontal direction, local solutions in a vertical direction must be known.

Is the scale-invariant, area-average Darcy-Richards parameterization now inappropriate for use in large scale hydrological and meteorological models? No. The simplicity of a single set of a large scale soil hydraulic parameters to represent the area-average, at the same time embracing the underlying soil heterogeneity, is at present too attractive to be discarded. The existence of a large-scale DR - solution for unsaturated soil moisture flow has neither been unambiguously proven and accepted, nor disproved and rejected. However, the lack of a consistent physical-mathematical proof restricts the use of the DR-equation at large scales to a convenient working hypothesis. Therefore the parameters describing the non-linear, area-average $h - \theta - K$ functions should be treated as *calibration-parameters*, which do not necessarily have a physical meaning. The advantages of this approach, when dealing with soil variability in the parameterization of soil water fluxes, will become apparent in some of the following sections.

3. Averaging of saturated hydraulic conductivity

The saturated vertical hydraulic conductivity K_s varies strongly between soil types (Clapp and Hornberger, 1978; Cosby et al, 1984; Wosten et al, 1995). In predictive models of soil hydraulic conductivity (Van Genuchten et al, 1992; Clapp and Hornberger, 1978 and Mualem, 1976, among others) K_s is generally treated as a

measurable parameter. K_s determines to a large extent the partitioning of rainfall between surface runoff and infiltration in most SVAT-models and therefore these models show a high sensitivity to the variability in K_s (Shao et al, 1994; Kabat et al, 1995).

The Darcy parameterization prescribes a linear relationship between the flux density and the hydraulic potential gradient. Unlike the unsaturated conductivity, the saturated hydraulic conductivity does not depend on flux density or on the soil water potential. In groundwater hydrology and petroleum engineering (Quintard and Whitaker, 1988; Stam, Zijl and Turner, 1989), it is concluded that the value of K_s strongly varies with the scale, but in general, the domain- average of K_s can be predicted from a known distribution of small scale K_s -values :

$$\langle K_s \rangle(X, Y, Z) = \iint \frac{K(x, y, z)}{LX LY} dx dy \quad (12)$$

$$K_{s, vol}(X, Y, Z) = \left[\int \frac{dz}{\langle K_s \rangle(X, Y, z) \cdot LZ} \right]^{-1} \quad (13)$$

where LX , LY and LZ are the dimensions of the 3 D domain centred around X , Y and Z . These equations show that the volume-average, large scale vertical hydraulic conductivity $K_{s, vol}$ equals the harmonic mean in the vertical direction (equation (13)) of the arithmetic mean of small scale conductivity in the horizontal direction $\langle K_s \rangle$ (equation (12)). It is interesting to note that in the case of K_s , the averaging in *both horizontal and vertical directions* can be done *without explicit a priori* knowledge of the small scale solutions of the soil water fluxes. In principle, Equations (12) and (13) can be used to determine vertical $\langle K_s \rangle$ of multi-layer profiles, consisting of different textural soil types. In the special case of a flow within a shallow, homogeneous single-layer soil profile, the change in water storage can be neglected and $\langle K_s \rangle(X, Y, z) \approx K_{s, vol}(X, Y, Z)$. In a case study presented later (section 6), this assumption has been tested and compared to aggregated K_s according to a method of Noilhan and Lacarrere (1995).

4. Scaling of soil characteristics

Scaling techniques reduce the effects of variability in soil parameters through scale factors. These are conversion factors that relate the characteristics of one system to the corresponding characteristics of another system (Tillotson and Nielsen, 1984, Hillel and Elrick, 1990). In contrast with averaging procedures, scaling tends to extrapolate parameterizations (relationships) rather than measurements. In general, this should be sufficient for large scale hydrological and meteorological models where areal fluxes need to be computed. In soil science, two main groups of methods are used to derive scale factors for the soil characteristics: functional normalization and dimensional methods (Kuhnel, 1989). In functional normalization, single-valued scale factors are derived for each experimentally established relationship through regression analysis (e.g. Russo and Bresler, 1980). Dimensional methods are based on the concept of similarity (Tillotson and Nielsen, 1984). In physical systems, three types of similarity can be defined: geometric,

kinematic and dynamic. In this paper, geometric and dynamic similarities will be discussed.

Soils as geometrically similar media

Geometric similarity in soils (Miller and Miller, 1956) is widely applied. The concept is based on the similarity of two soils which are geometrically alike except for the scale of their internal porosities. Figure 1 shows two geometrically similar soils, which have geometrically similar distributions of air and water in their respective pore spaces. Assuming geometrically similar media, the variability in $h - \theta - K$ characteristics can be described by single scale factor α . At a certain location x within the flow domain this scale factor can be expressed as:

$$\alpha_x = \frac{\lambda_x}{\lambda_r} \quad (14)$$

where λ_x and λ_r are the characteristic length scales respectively of a soil at location x and of a reference soil. From viscous flow theory (Raats, 1990) the following scaling rules can be derived for the soil moisture content θ , pressure head h and hydraulic conductivity K :

$$\theta_x = \theta_r \quad (15)$$

$$h_x = \alpha_x^{-1} h_r \quad (16)$$

$$K_x = \alpha_x^2 K_r \quad (17)$$

These equations imply that the soil water retention and hydraulic conductivity characteristics at a given water content θ_x at any location can be related to mean $h - \theta - K$ functions, and are of a general form:

$$\zeta_x = \alpha_x^n \zeta_r \quad (18)$$

Parameterization schemes for soil water fluxes vary widely in the SVAT models used in large scale models and in GCMs (Shao et al, 1994). For many variables in soil parameterizations common in SVAT schemes, so called secondary scaling rules can be inferred from the form of the Darcy equation and the mass conservation equation (Raats, 1983, 1990). Values of n in equation (18) are given in Table 1.

Raats (1990) noted that the log-normal distribution of the characteristic lengths λ (Equation 14) implies that the scaled variables in Table 1 will also be log-normally distributed. This is in agreement with investigations of Nielsen (1973), Peck et al (1977), Sharma et al (1980) and Ragab and Cooper (1993 a,b) who reported log-normal probability distribution functions of hydraulic conductivity, water diffusivity and sorptivity from field studies. The following example of similar - media scaling

Table 1

Values of exponent n in scaling equation (18) for several variables and parameters used to model soil water flow. D is the water diffusivity ($D = k \, dh/d\theta$), MFP is a matrix flux potential (Shaykewich and Stroosnijder, 1977) and v is the velocity; for other variables see text.

variable ζ :	θ	h	H	D	MFP	v	q	x,y,z	t
exponent n :	0	-1	-1	1	1	2	2	-1	-3

deals with hydraulic characterisation of sandy soils during the Hapex-Sahel experiment (Soet et al. 1993, Kabat et al., 1995b). Using sampling schemes as given in Figures 2 a,b, a population of undisturbed soil samples was collected at a depth between 10 and 15 cm. Hydraulic characteristics of these were determined in the laboratory (van Dam et al, 1994) and are given in figures 3 a,b. The spatial variability is mainly evident in the dry part of both curves, below a moisture content of about 15 %. Also, the near-saturated hydraulic conductivity varies within the population from 1 to about 60 cm/hour. Simultaneous similar media scaling (Clausnitzer et al, 1992) was applied to all functions, resulting in *reference curves* for soil water retention and hydraulic conductivity. In Figures 4a,b these reference curves are shown together with optimized curves, which resulted from "true-averaging" of the outflow in a multi-step determination procedure (van Dam et al, 1994). The similarity of the 2 curves suggests that *the reference curve, obtained through scaling, does closely approach the "effective", area-average curve*. Indeed, an independent calibration of a 1-dimensional Soil-Vegetation-Atmosphere-Transfer model SWACROP/SWATRE (Kabat et al, 1992; Feddes et al, 1978) with a sophisticated DR-parameterization of the soil water flow, using measured soil water data during the Hapex-Sahel experiment, showed the best fit with the h - θ - K curves which were almost identical to the reference curves (Meeuwissen, 1995). Similar conclusions about the "effective" meaning of similar - media reference h - θ - K curves will be reached in section 5 of this paper.

Dynamic scaling methods

It has been argued by several authors that natural soils do not always meet the criteria of similar media (Sposito and Jury, 1985; Jury et al, 1987). This has led to the development of dynamic scaling methods, such as inspectional analysis. Inspectional analysis involves a simplification process by which the number of dimensional quantities describing a physical system is reduced to a smaller number of non-dimensional quantities. Replacing the $K(h)$ function by the water diffusivity D Equation (1) can be written as:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[D \frac{\partial \theta}{\partial z} \right] - \frac{\partial K}{\partial z} \quad (19)$$

The water diffusivity D in eq. 19 is defined as $D = -K \, \delta h / \delta \theta$. The time and depth

variables can be converted into dimensionless form (Kuhnel, 1989):

$$\tau = \frac{K_s^2 t}{D_0} \quad \xi = \frac{K_s z}{D_0} \quad (20 \text{ a,b})$$

where D_0 is the water diffusivity at air-dry soil moisture content. Equation 19 can now be written in dimensionless form :

$$\frac{\delta \theta}{\delta \tau} = \frac{\delta}{\delta \xi} \left[D/D_0 \frac{\delta \theta}{\delta \xi} \right] - \frac{d(K/K_s)}{d\theta} \frac{\delta \theta}{\delta \xi} \quad (21)$$

Following inspectional analysis, two soils have similar concentration versus depth or time curves if they have identical values of τ and ξ . According to Kuhnel (1989), two universal scale factors can be derived for both time and depth dependent soil water flow :

$$\delta = \frac{(D/K_s^2)_x}{(D/K_s^2)}, \quad \varepsilon = \frac{(D/K_s)_x}{(D/K_s)}, \quad (22 \text{ a,b})$$

where the subscripts denote respectively the "similar" soil at any location x and the reference soil.

This concept can be applied when dealing with sub-grid variability of soil moisture fluxes in mesoscale and large scale models. For the large scale flow domains, $z \sim 0$ and therefore $\xi = 0$. This means that only one scaling parameter is required for the scaling of (dimensionless) flow processes close to the land-surface atmosphere interface, such as infiltration and soil controlled evaporation. It is very attractive that a single characteristic in the form of D/K_s^2 can be used to quantify the response of similar soils in terms of their hydraulic behaviour. Values of D/K_s^2 vary with textural soil types. In figure 5, characteristic D/K_s^2 times, as derived in Kuhnel (1989) are given for a full range of textural soil types, as compiled by Mualem (1976) and using the US Soil Conservation Service soil classification. Diffusivity was kept constant at $\theta = 0.3$.

One of the implications of this type of soil classification is that if the sub-areas with the same order of magnitude of D/K_s^2 can be identified on regional soil maps, one can allocate areas, compatible with the subgrid scale of mesoscale models, which contain similar soils (in terms of comparable hydrological response). To compute absolute soil moisture fluxes and their spatial distribution within a subgrid, spatial distributions of δ and K_s must be known. Analysing the results of Nielsen et al (1973), who found a log-normal distribution of a scaling parameter for K_s in the context of geometrically similar media, Kuhnel (1989) concludes that δ and ε are related to the scale factor α through $\delta = \alpha^2$ and $\varepsilon = \alpha$. Since α has a log-normal distribution (see also above), $\delta^{1/2}$ will have a log-normal distribution as well (Raats, 1983). This means that the scaling of soil fluxes within a sub-grid only involves the choice of reference values of D/K_s^2 and K_s for the sub-grid area, where the spatial variability of these parameters can be determined as deviations from $\delta = 1$ and $\delta/\varepsilon = 1$.

There are several closed analytical solution for water diffusivity, such as (Fujita 1952):

$$D = \frac{D_0}{(1 - s\theta)^2} \quad (23)$$

where D_0 is a fixed value for each textural soil type and s is the slope of the soil moisture retention curve. In analogy with $\langle K \rangle \langle X, t \rangle$ (see Equation 8), the large-scale, area-average diffusivity can be expressed as:

$$\langle D \rangle \langle X, t \rangle = \langle D(\theta(x, t)) \rangle \langle X, t \rangle \quad (24)$$

and s can be determined for each value of θ . Therefore, knowing the area average $\langle K_s \rangle$ from Equations 12 and 13, a simple arithmetic averaging can be applied to arrive at area-average, effective diffusion time $\langle D / K_s^2 \rangle$. For a fixed value of the volumetric soil moisture content, the heterogeneity within a sub-grid consisting of several textural soil types can be therefore described by a single "effective" scaling parameter $\langle D / K_s^2 \rangle$.

5. Effective soil parameters from inverse modelling

Inverse modelling has been widely applied in groundwater hydrology (e.g. Wagner and Gorelick, 1986;) and since the early eighties, also in unsaturated soil water flow problems (e.g. Mantoglou and Gelhar, 1987b; Kool et al, 1987; Mishra et al, 1990). For the latter, the studies have been limited to calibration experiments with laboratory-size core soil samples or with lysimeter data (van Dam et al 1994, Kool et al, 1987, among others). In this way the effective soil hydraulic properties integrate micro-scale variability, limited to the size of a soil core. Using these parameters for large flow domains, the problem of spatial heterogeneity remains.

Feddes et al (1993a, b) suggested how to estimate the effective soil hydraulic functions for unsaturated soil water flow from measurements of entire-domain hydrological responses. There is some novelty in this idea, as it envisages the use of a new generation of scale-integrating measurements of the surface soil moisture (Schmugge et al, 1992), areal evaporation fluxes (Shuttleworth, 1991) and remote-sensing driven techniques for estimation of the areal evaporation fluxes (Bastiaanssen et al, 1994). Several studies since reported application of this technique with field data (Hollenbeck, 1992; Rainin et al, 1995,). The study by Feddes et al (1993b) will be used to demonstrate the potential of this technique to describe large scale parameterization of the soil hydraulic behaviour.

For a 6.5 km² catchment (figure 6) distributed soil samples were taken and the soil hydraulic properties were determined in the laboratory by Hopmans and Stricker (1989). The method of Warrick et al (1977) and Russo and Bresler (1980), was used to determine similar - media reference curves for the water retention and the hydraulic conductivity, a log-normal distribution of the scaling factors and a normal distribution of the water content at saturation. The reference curves were described by the Mualem-van Genuchten (MVG) predictive model (1980):

$$S = [1 + (\alpha|h|)^{\eta}]^{-m} \quad (25)$$

$$K = K_s S \left[1 - (1 - S^{1/m})^m \right]^2 \quad (26)$$

where $S = (\theta - \theta_r) / (\theta_s - \theta_r)$. The subscripts r and s denote residual and saturated volumetric water content, while n , m , α , and l are shape parameters of the MVG-model. Including K_s , θ_r , and θ_s , the MVG predictive model is therefore determined by 6 coefficients ($m = 1 - 1/n$). These coefficients are the calibration parameters of effective soil water retention and hydraulic conductivity curves. In this study, reference curves together with distributions of the scaling factors were used to randomly generate the soil hydraulic properties of 32 hypothetical bare soil columns. In tables 2a and 2b, the characteristics of the generated spatially variable soils are given.

Table 2a

MVG coefficients of the similar-media reference curve used in a large-scale inverse parameterization of the DR-equation.

variable:	α	n	K_s	θ_r
dimension:	cm^{-1}	-	$\text{cm} \cdot \text{d}^{-1}$	-
coefficient:	0.01924	1.59	33.7	0

Table 2b

Statistical properties of the similar-media reference curve used in a large-scale inverse parameterization of the DR-equation.

distribution type:	log normal $^{10}\text{Log}(\alpha)$	normal θ_s
mean:	-0.1155	0.4060
standard deviation:	0.3430	0.0354

Simulations with a SWACROP/SWATRE model (Feddes et al, 1978; Kabat et al, 1992) were executed for each of the 32 bare soil columns for different sets of meteorological forcing data and initial conditions. These sets were chosen to represent a reasonably wide range of flow situations (wet, dry, transient). The water fluxes and water content in the profiles were then averaged to obtain area-average, integrated values (figure 6). The inverted numerical model, applied to the entire catchment represented by one "effective" column, was then optimized such that the results closely match the area-average value. Different combinations of area-average hydrological variables were used in the

optimization procedure:

- i) Volumetric water content of the profiles on selected days;
- ii) Cumulative evaporation and infiltration data. This set was selected anticipating the availability of areal-evaporation data from direct measurements by aircraft, for instance in several large scale land surface - atmosphere experiments or from remote sensing algorithms (Bastiaanssen et al, 1994)
- iii) Cumulative evaporation and infiltration together with the surface volumetric moisture content. In addition to the variables in ii), surface volumetric moisture content was selected to show the potential use of integrated, areal-estimates of surface soil moisture by passive and active microwave techniques (Schmugge et al, 1992)

Two different optimization procedures were used: a Price-algorithm (Price, 1979) and a Levenberg-Marquard method (Marquard, 1963). For more details, the reader is referred to Feddes et al (1993b) and de Rooij and Schuitema (1990). The resulting effective, area-average soil hydraulic curves are shown in Figures 7 a, b, c, d. The numbering of individual curves refers to the following combinations of calibration data sets: (1) all 6 parameters of MVG model calibrated on soil moisture profiles; (2) as 1, but θ_r and θ_s fixed; (3) all 6 MVG-parameters calibrated on cumulative evaporation/infiltration fluxes; (4) all 6 MVG-parameters calibrated on cumulative evaporation/infiltration fluxes, supplemented by θ at 0.5 cm depth; (5) as 4, but θ_r and θ_s fixed. For comparison the reference curves obtained through similar scaling are shown (6). Figure 7 suggests relatively large discrepancies between individually generated effective curves with the exception of the curves (1), (2) and (5). The similarity of the two effective curves based on the optimisation with soil moisture profile data (1 and 2) with the scaled-reference curve is striking. Curve (4) shows the largest deviation from the remaining curves, suggesting that the addition of surface soil moisture data into the objective function did not improve the optimisation results. In fact, the opposite happened. This can probably be attributed to the inaccuracy in the definition of the error covariance matrix and associated weight coefficients in the two-component (i.e. evaporation/infiltration flux and volumetric moisture content) objective function (Kool et al, 1987).

A verification experiment tested the response of the effective flow domain through simulations of the seasonal water balance of the catchment, covered with grass. A dry (1976) and a wet (1982) year were selected, so meteorological forcing data, initial conditions and the boundary conditions were outside the range over which the effective parameters were optimized. The simulations were again performed with the SWATRE/SWACROP model. First, simulations were run for the 32 soil columns, each with different soil hydraulic characteristics (see tables 2a, 2b) and covered with grass. The resulting water fluxes and profile moisture contents were then arithmetically averaged to create a *synthesised area-average response*. Single runs for a single, effective column were then performed with the effective soil hydraulic parameters obtained from the calibration with soil moisture profiles

(1), evaporation fluxes (3) and with the similar media-reference curve (6).

In Figures 8 a - h, results are presented for selected days within the season. In both dry and wet years, the area-average soil water content at top and at the bottom of the soil profile (Figures 8a-d) was predicted by all effective curves with sufficient accuracy for most of the days (the maximum standard deviation of the 32 collum mean was around 0.07). The same conclusion can be drawn for predicted cumulative actual transpiration (Figures 8e-f). The superior performance of the effective curve (1), as well as very acceptable results obtained with the similar media reference curve are both noteworthy. The cumulative discharge to groundwater (Figures 8g-h) was slightly underestimated in all cases. The effective curves obtained from fitting to evaporation fluxes did not adequately predict area-average discharge. Again, the reference, scaled curve performed as well as the best set (1).

These results are encouraging. First, they show that area-average, effective parameterization of the soil hydraulic functions in combination with a numerical solution of the scale-invariant DR-equation can work well in some cases. The reference soil hydraulic functions obtained through geometric scaling performed well in all cases; in accordance with a similar conclusion drawn for sandy soils in the Sahel (section 4). This is an indication that the reference curves can be used as effective curves representing the area.

The choice of the variables to test the performance of inversely derived, area-average soil hydraulic parameters was not arbitrary. The moisture content close to the surface (in this case at the depth of 1.5 cm) largely influences surface runoff generation in large scale models. The moisture content close to the bottom profile boundary has a similar importance for modelling of the subsurface runoff (deep percolation). Both variables were predicted well by large-scale effective soil hydraulic parameters. It can be concluded that use of these parameters in soil hydrological parameterization schemes of large-scale atmospheric models should improve the performance of these schemes in a hydrological sense, i.e. when predicting seasonal cycles of the soil moisture and other terms of the water balance. Cumulative transpiration was predicted well by all effective parameter sets. This is attributed to the averaging power of the vegetation root water uptake activity. However, the effective curves obtained from fitting to the evaporation fluxes failed to predict the groundwater recharge with sufficient accuracy. This points towards problems when the same "vegetation averaging power" is included in the optimization without additional data on soil profile moisture. The (negative) change in water storage of the root zone is mainly caused by the root water uptake (see Equation 1) and therefore the other hydrological fluxes are less strongly driven by direct changes in soil moisture potential due to the Darcian water flux.

The effective parameters derived by several optimization data sets and with independent similar media technique are almost identical, which means that they represent a unique set for this area. Under normal conditions, these parameters are also time-invariant.

6. Aggregated soils

Techniques to estimate representative, aggregated soil hydraulic characteristics of

sub-grids with mixed soil types are subject of the discussion in this section.

A pedological (textural) soil type is defined in this context as a soil block having a characteristic particle size distribution (texture) and related physical and chemical characteristics. These soil types (sand, clay, loamy sand etc) are usually assigned a typical set of soil water retention characteristics and hydraulic conductivity curves. For further discussion it is essential to underline that the concept of discrete soil types is a simplification adopted by soil scientists for practical purposes such as soil mapping. The classification of the pedosphere continuum into discrete types is unavoidably arbitrary at the limits of the soil type (Scholes et al, 1995). Soil profile descriptions are samples out of a distribution of possible values. Therefore the properties of a soil type are described in a statistical sense. Consequently, soil map units do not represent homogeneous soil types, but associations of given proportions of several soil types. Knowing this restriction in the definition of soil types, it is theoretically possible to define a "soil functional ensemble": this is a mixture of "discrete" soil types with a common reference through an arbitrarily chosen, integrated hydrological parameter. This could be for example the range of available water storage between field capacity and wilting point (Kabat and Beekma, 1994). Common examples of "soil functional ensembles" are light, medium and heavy soils.

Typical sub-grids of mesoscale atmospheric models would usually cover several such pedological soil types. Due to strong non-linearity in the relationship between soil texture and soil hydraulic characteristics, the aggregation of soil texture classes presents a non-trivial task. Aggregated soils should, like "effective" soils, result in the same dynamic response of the hydrological system, compared to an area-average obtained by procedures which retain the soil variability.

Cosby et al (1984) showed that the soil hydraulic characteristics of 11 textural classes included in the USDA-soil classification system (Figure 5) were mainly determined by the particle size distribution. The mean values and variances of the parameters of the Clapp and Hornberger (1978) exponential approximation of the soil hydraulic characteristics (Equations 27-29) were found to be linearly dependent on the soil texture. Using multilinear regression analysis, Cosby et al (1984) also found that most of the variability in soil parameters can be related to the clay and sand fractional contents. Assuming the pedosphere to be a continuum with continuous variations in the soil properties related to the variation in the sand and clay content, Noilhan and Lacarrere (1995) derived aggregated soil hydraulic parameters for an area comparable in size with a GCM-grid in HAPEX-Mobilhy. Using both one and full three dimensional numerical atmospheric modelling, they found that, despite the non-linear relationship between evaporation and the soil hydraulic characteristics, the area-average surface fluxes computed with effective soil parameters by the 1-d model agreed to within 10 % with the area-average fluxes calculated by the 3-D model. On the other hand, using area-dominant soil characteristics resulted in a significant departure of the predicted fluxes from the true area average.

A new simulation experiment has been set up to test the overall performance of the soil aggregation procedure introduced by Cosby et al (1984) and Noilhan and Lacarrere (1995). A surface grid consisting of 3 textural soil classes has been submitted to an aggregation procedure. The soil types are clay with a fractional

area equal to 4/9, sand having the fractional area of 3/9 and the loam with fractional area equal to 2/9. Soil hydraulic characteristics are described by the Clapp and Hornberger (CH) (1978) model:

$$h(\theta) = h_s(\theta/\theta_s)^{-b} \quad \theta < \theta_p \quad (27)$$

$$h(\theta) = -m(\theta/\theta_s) - n(\theta/\theta_s - 1) \quad \theta_p \leq \theta \leq \theta_s \quad (28)$$

$$K(\theta) = K_s \left(\frac{\theta}{\theta_s} \right)^{2b+3} \quad (29)$$

where θ_p is the moisture content corresponding to the inflection point of the soil water retention and hydraulic conductivity curves. The values of the parameters of the CH model for each of the soils were selected according to Cosby et al (1984) and are given in table 3.

Table 3

Clapp and Hornberger parameters of the soils used in the soil aggregation case study.

parameter: dimension:	θ_s	h_s m	b	θ_p	K_s m.s ⁻¹
Sand:	0.44	0.0729	3	0.39	$2.5 \cdot 10^{-5}$
Loam:	0.48	0.1476	5	0.432	$7.0 \cdot 10^{-6}$
Clay:	0.52	0.2152	8	0.468	$2.0 \cdot 10^{-6}$

To each of the soil types, mean clay and sand percent compositions (C_i and S_i) have been assigned according to Clapp and Hornberger (1978). For example, the textural soil type "clay" has a mean clay fraction equal to 0.63, and a sand fraction of only 0.03.

The area aggregated texture was then estimated from the respective clay and sand fractions as an arithmetic weighted average:

$$C_{aggr} = \sum_i f_i C_i \quad S_{aggr} = \sum_i f_i S_i \quad (30)$$

Using the regression between the clay fraction and the b-parameter in the CH model and between the sand fraction and θ_s , as proposed by Noilhan and Lacarrere (1995), aggregated values of b and θ_s were derived. To preserve the non-linear relationship between saturated hydraulic conductivity and the clay content, the aggregated K_s value was derived for the corresponding area-aggregate clay fraction from the table in Clapp and Hornberger (1978). In addition and with the aim to test effects of this non-linearity, an alternative linear averaging of K_s according to

equation 12 has also been performed.

The resulting aggregate soil water retention and hydraulic conductivity curves are shown in figs 9a, b, together with the curves of the three individual soil types. Linear averaging of K_s yields an aggregate K_s (Aggr-avg in Figure 9b) which is in absolute terms about half the aggregate K_s , derived from the aggregate clay fraction from the CH table (Aggr-LUT in Figure 9b). The general non-linearity inherent to this aggregation procedure is demonstrated by the K -values in the dry part of the curves ($\log |h| > 3$). Both aggregated K -curves lay above the K -curves for individual soils, therefore giving higher values of the aggregated conductivity compared to any of the individual soils. This can be attributed to the aggregated b -parameter, which was higher ($b_{\text{aggr}} = 8.1$) than the b -values for each of the individual soil types (Table 3).

A 1-dimensional SVAT model SWAP/MITRE (Kabat et al, 1995c; Dolman, 1993) has been used to assess the effects of soil aggregation on different terms of the water balance in the grid. SWAP/MITRE combines a fully dynamic, DR-parameterization of the soil water flow (Feddes et al, 1978; Kabat et al, 1992) with a multilayer surface energy balance model (Dolman, 1993). A single-layer, 2 meter soil profile without presence of groundwater has been assigned to the tested grid. Meteorological forcing and vegetation parameters (grass) have been obtained from a pasture site of the Rondonia Boundary Layer Experiment (RBLE 3) in Brazil and were kept constant for all simulations. Simulations have been executed for a consecutive period of 12 days with imposed total rainfall of 97 mm, distributed between several showers with an average intensity not exceeding 5 mm/h.

The results are presented in Figure 10. To demonstrate the hydrological behaviour of each of the individual soil types, water balance terms are first plotted separately for each soil type (Figs 10 a, b). Cumulative vegetation evaporation on the clay soil (14 mm) is reduced to less than one third of the evaporation on the sandy soil (45 mm), while the cumulative bare soil evaporation of all three soil types remains about the same (~ 6 mm). This can be explained when inspecting the actual shape of the water retention curves (Figure 9a) in combination with the soil profile moisture. On average, the volumetric soil moisture content over the root zone was about 0.2. For sand, this corresponds to a soil suction of ~ 450 cm, while the same volumetric moisture content in clay soil is related to a soil suction of $\sim 10^5$ cm ! The water in sandy soils is therefore still easily extractable for vegetation, while the high suction of the clay soil effectively prevents the root water uptake. The remaining terms of the water balance are shown in figure 10b.

There are two conclusions to be drawn at this point. First, if the domain tested in the present case study was aggregated to the dominant soil type (clay) within a grid, a severe underestimation of area-average evaporation would occur. This is in full agreement with the results of Noilhan and Lacarrere (1995). Second, the deep percolation flux (subsurface runoff) differs widely between the 3 soil types which means that a dominant-soil approach within a grid would also result in a substantial underestimation of the area-average subsurface runoff.

Fully distributed runs were subsequently used to calculate true-average water balance terms (Areal in Figures 10 c-d), against which the results with aggregated parameters were tested. Figure 10c shows cumulative areal-evaporation calculated with both aggregated parameter sets together with the true average evaporation. The aggregated parameters predict the total area evaporation

with a high accuracy of about 10 %. This might be surprising when looking at the aggregated water retention curve (Figure 9a). However, the area-average, moisture content in the root zone was close to $0.3 \text{ m}^3\text{m}^{-3}$ and well within extractable ranges of the aggregated soil suction. In Figure 10d, subsurface runoff and the total change in profile water storage are shown. The true area average subsurface runoff exceeds 100 mm and the area-average water change in profile water storage was -35 mm. These values are mainly influenced by the low water holding capacity of the sand. The aggregated soil parameters drastically underestimated subsurface runoff; by more than 80 %. The error in total profile water storage was around 40 %.

It is concluded that aggregated soil parameters, obtained by the aggregation procedure introduced by Cosby et al (1984) and Noilhan and Lacarriere (1995) may be used to predict the area-average evaporation flux. This in agreement with previous conclusions of Noilhan and Lacarriere (1995) and indirectly, with Braud et al (1995), who stressed the averaging-power of the vegetation, smoothing out the effects of variability in soil properties.

However, the aggregated parameters fail in predicting the water balance terms associated in a more direct way with the soil water flow processes, such as the downward percolation flux or runoff. This is a serious drawback that is increasingly problematic as the time span of model integrations increases to seasonal, multi-annual or even longer time scales.

7. Discussion and conclusions

A large spectrum of the techniques to parameterize and scale up heterogeneous soil hydrological processes exists. The motivation for the development of these techniques originates primarily from the need of the two basic characteristics of the soil water flow: the soil water retention and hydraulic conductivity curves. These curves show a strong non-linear behaviour, which complicates the scaling procedures. At the large scales, such as the gridcells of atmospheric mesoscale and climate models, the problem is exacerbated by questions about the scale-invariance of the basic parameterization schemes of the soil water flow, such as the Darcy-Richards equation.

Different relationships have been proposed and tested to describe the non-linear relationship between soil water content, soil suction and the hydraulic conductivity. In this paper, the two most widely used schemes have been presented and used in numerical experiments: the MVG- conceptual model and the CH-conceptual model. Optimization and/or regression procedures are used to fit the data to these pre-defined conceptual models. If this procedure is applied at large scales, the fitting parameters must be considered as pure algebraic artefact without direct physical meaning. However, their values are submitted to constraints imposed by the use of the basic transfer equations.

It has been shown that the use of these soil characteristic conceptual models, combined with appropriate initial and boundary conditions allow the description of water transfer in the unsaturated zone of the soil in a fully deterministic way. Consequently, one can derive the area-effective soil hydraulic

parameters. However, such a comprehensive approach pays its price in complexity. Therefore another direction is developing in the field of soil science and hydrology, that is potentially interesting for large-scale modelling: simplification of the $h - \theta - K$ conceptual models, applicable to the field data. The objective is to minimize the number of parameters needed to describe the non-linear relationship between soil moisture, pressure and conductivity and at the same time to retain much of the dynamics of soil water fluxes.

When dealing with large areas, most of the field methods to collect soil hydraulic data are becoming too cumbersome to apply on a routine basis. In the last three decades, this has led to many attempts to predict soil hydraulic characteristics from more easily accessible soil data, such as soil textural information (e.g. Van Genuchten et al, 1992, Wosten, 1995). One technique, which was not discussed in this paper, introduces so called pedotransfer functions which are used to average the soil hydraulic properties from known area-distributions of the soil (textural) properties (e.g. Wosten et al 1995). In many ways, these techniques resemble the example of soil aggregation given in section 7 of this paper.

Except when accounting for the soil heterogeneity by introducing stochastic soil water models into coupled meteorological-hydrological models (e.g. Chen et al, 1994), the aggregation and scaling of soil properties seems largely to be an *off-line* task, which does not necessarily enter the actual calculation procedure of a mesoscale model. This contrasts with patch scale aggregation for mesoscale meteorological modelling, which in some cases depends on interactions between surface parameters and the overlying meteorology (Dolman & Blyth, this issue). The (theoretically) conservative behaviour of soil hydraulic parameters should be seen as a fortunate property which simplifies the implementation of area-effective soil parameters into large scale atmospheric and climate models. At this point, another consequence emerges regarding the *off-line* approach to the effective parameterization and aggregation of the soil hydraulic properties at the spatial scales compatible with grid-size of atmospheric and climate models. This concerns the availability of *compatible basic soil data* which are needed to apply the aggregation and scaling rules.

The three major sources of soil data - the FAO, USDA-SCS and ISRIC (Scholes et al, 1995) - use different soil classes and classification criteria. Harmonisation of these would strongly facilitate the use of these data in climate modelling. National soil data bases, traditionally sufficient for local and regional hydrological studies, have shown to be inconsistent in the context of global modelling. Climate modellers need not only average values of water holding capacity or bulk density, but also a spatial distribution of these characteristics. Traditional techniques like kriging are not applicable in this context, because the distances between different soil profiles within the grid of a climate model are usually many orders of magnitude larger than the spatial scale of auto - correlation. Clearly, these issues need to be tackled in a close interaction between the soil science community and the meteorological and climate modelling community. Current cooperative efforts between GEWEX-ISLSCP and IGBP-BAHC projects in the framework of the so called Initiative II Global Data Sets are one of the first steps in this direction.

The techniques presented in this paper to scale and aggregate the soil characteristics all potentially qualify for use in large scale meteorological models. One of the most attractive and perhaps surprising results is the effective behaviour of the reference curve, obtained through similar media scaling (section 4). Further studies are needed, but if this conclusion holds, a relatively simple method will become available to parameterize soil variability at large scales. The inverse technique was found to provide *effective soil parameters* which performed well in predicting *both* the area-average evaporation and the area-average soil moisture fluxes, such the subsurface runoff. This was not the case for *aggregated soil parameters*, obtained from a regression type of relationships between soil textural composition and the hydraulic characteristics. The aggregated parameters, although predicting the evaporation fluxes well, failed to predict the water balance terms, such as downward percolation flux or runoff, that are associated in a more direct way with the soil water flow processes. This is a serious drawback which could eventually hamper the efforts focused on improvement of the hydrological cycle and overall parameterization of soil hydrological processes in mesoscale atmospheric models and in GCMs (Shao et al, 1994).

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Figure 10.

Results of the SWAP/MITRE simulations with aggregated and separate soil hydraulic characteristics as in Figure 9. (a) Cumulative vegetation transpiration (Ea) and soil evaporation (Es) for individual soil types. (b) Cumulative subsurface runoff (=deep percolation, D) and the change in profile water storage (dS) for individual soil types. (c) True area-average transpiration (EaAreal), soil evaporation (EsAreal), and the areal evaporation obtained with aggregated soil hydraulic characteristics (EaAgg). (d) Idem for subsurface runoff (D) and the change in profile water storage (dS).

Answers to questions

Q 1 (dr. B. Choudhury):

When doing the spatial averaging of the Richards equation, do you get a co-variance term with respect to water diffusivity?

Answer: yes, one gets a co-variance term because the water diffusivity D is a function of the volumetric moisture content θ , which varies spatially. The same is true when averaging the hydraulic conductivity rather than diffusivity. However, the non-linearity in the $D(\theta)$ function compared to non-linearity in the $K(\theta)$ function is smaller (D is better defined in a dry region compared to K). So the assumption of constant diffusivity for a large range of the moisture content is acceptable.

Q2 (dr. J. Schaake):

The examples you have used apply to motion of water in the vertical direction. In natural catchments, three-dimensional movement of water occurs and some aspects of this are central to understanding the water budget at large scales. The horizontal movement occurs only over distances ranging from tens to hundreds of meters. Additional parameters are topographic factors such as hillslope slopes and lengths. In what ways do the ideas you presented apply to the three-dimensional problem?

Answer: I believe there are two positive answers to your question. In a strict theoretical-mathematical sense and in a similar way to the Darcy-Richard's equation written as a scale-invariant (mathematical) expression, one can define mathematical expressions for full three dimensional soil water flow systems. A strongly simplified example is the averaging of saturated hydraulic conductivity in both horizontal and vertical directions. Techniques like similar media scaling or inspectional analysis can be applied in all three dimensions.

As discussed, spatial averaging of, for example, water retention curves is only possible with *a priori* knowledge of local scale solutions for soil water potential. A fully three dimensional, large scale unsaturated flow equation, accounting for co-variance terms in all three directions, leads to virtually unmanageable algebra. In cases with dominant lateral flow components, other techniques, like for example the well known TOP-model, might be therefore more useful.

One can also look for simplifications, i.e. for situations when a number of variables and dimensions can be ignored. For example: In a catchment somewhere in a region with tropical precipitation regimes, strong lateral flow (runoff, overland flow, interflow) would usually occur during the rainy season. However, at the end of the rains and during the dry-down period, vertical fluxes dominate and the system eventually becomes one-dimensional. The wet season hydrological regime data could be used for parameterization of the lateral flow processes, partly neglecting the vertical transport, while the dry-down period could be useful for parameterization of the vertical processes. I believe we should be looking for such "split-window" simplifications rather than trying to construct a full three-dimensional parameterization system which later needs to be simplified anyway because we can't find mathematical solutions to it. The soil hydraulic parameters are a unique set of functions which depend mainly on soil textural properties. They should not

depend on the flow, or whether one-dimensional or three-dimensional representations are used.

Q3: (prof. Fiedler)

I am not sure whether I understood your philosophy: what you called averaging seems to be the process of transferring the characteristics of soil media into appropriate physical parameters. When these are determined, we have a multiscale problem and we should be able, using asymptotic similarity theory, to reduce the necessary independent dimensionless variables to derive approximated equations depending on what scale we want to use these equations.

Answer:

The scale-dependent parameterizations (equations) are a theoretical possibility. The problem is the inherent continuity of the pedosphere as a medium and the continuous correlation lengths in this medium. However, for dimensionless scaling, for example the inspectional analysis by D/K_s^2 method introduced by Kuhnelt (1989) and discussed in this paper, some very promising approaches are emerging.

Q4 (dr. Detlef-Schultze):

In averaging soil water you cannot neglect plant cover, because it will determine porosity and the amount of deep drainage. Grasses and shrubs have different root structures and determine soil properties independent of soil types, like clay, sand and loam.

Answer:

I think you are referring to the macroporosities and flow paths in the soil caused by its structural changes. Indeed, roots will create preferential flow paths in the soil. When this preferential and or by-pass flow prevails above the matrix flow, the Darcy-Richards parameterization is not valid any more. In fact, this is one of the largest arguments for questioning the classical Darcy-Richards approach (Beven, 1994). However, there is a range of models which extend the DR-approach just for these effects.

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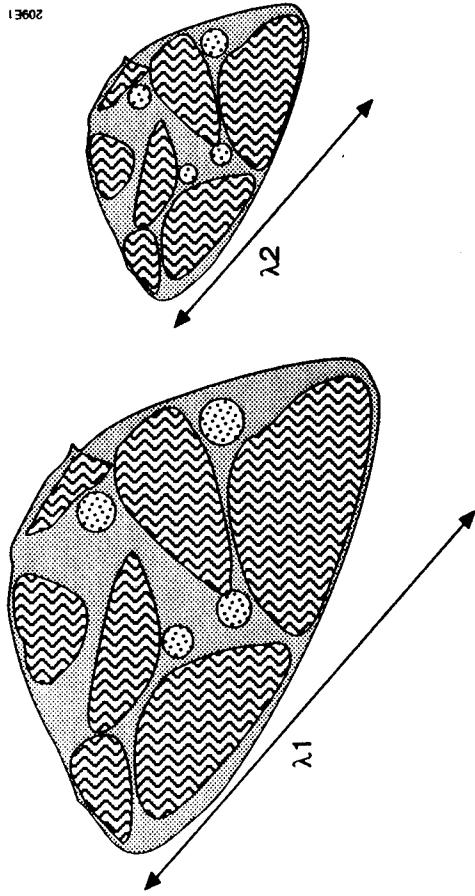


Fig 1

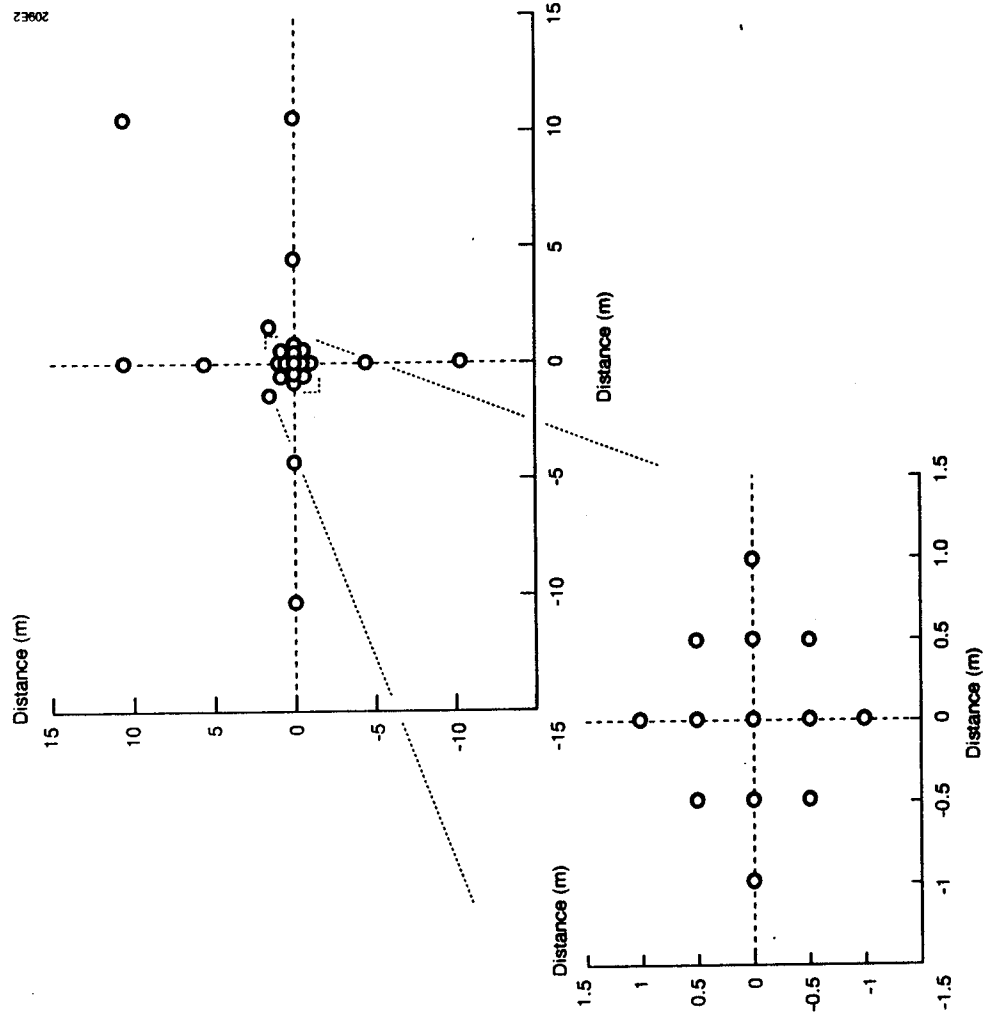


Fig 2

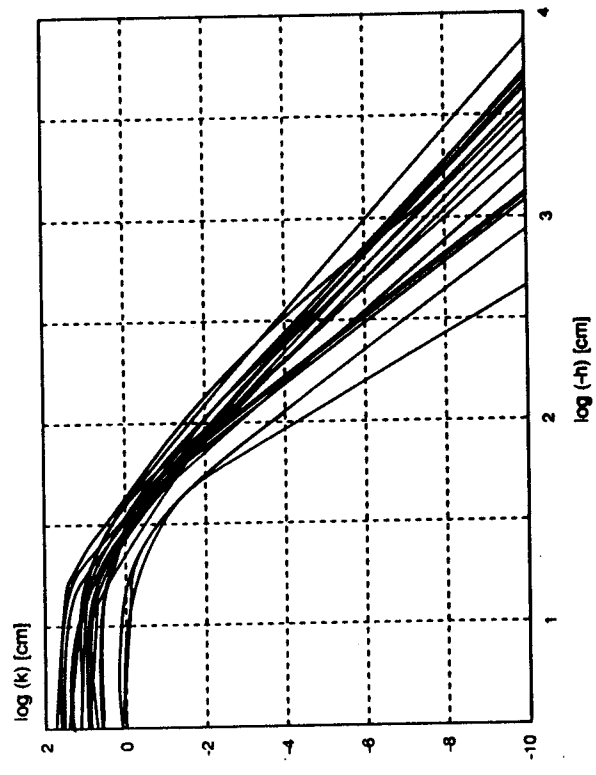
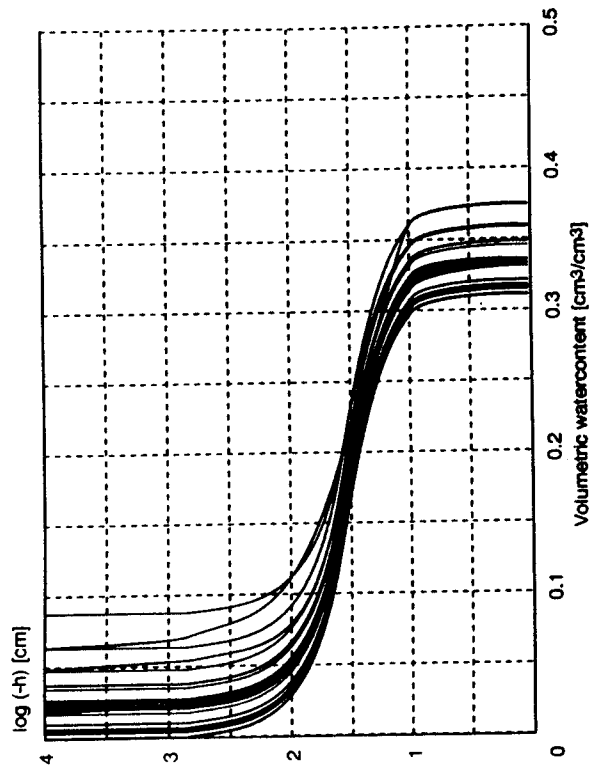


Fig 2

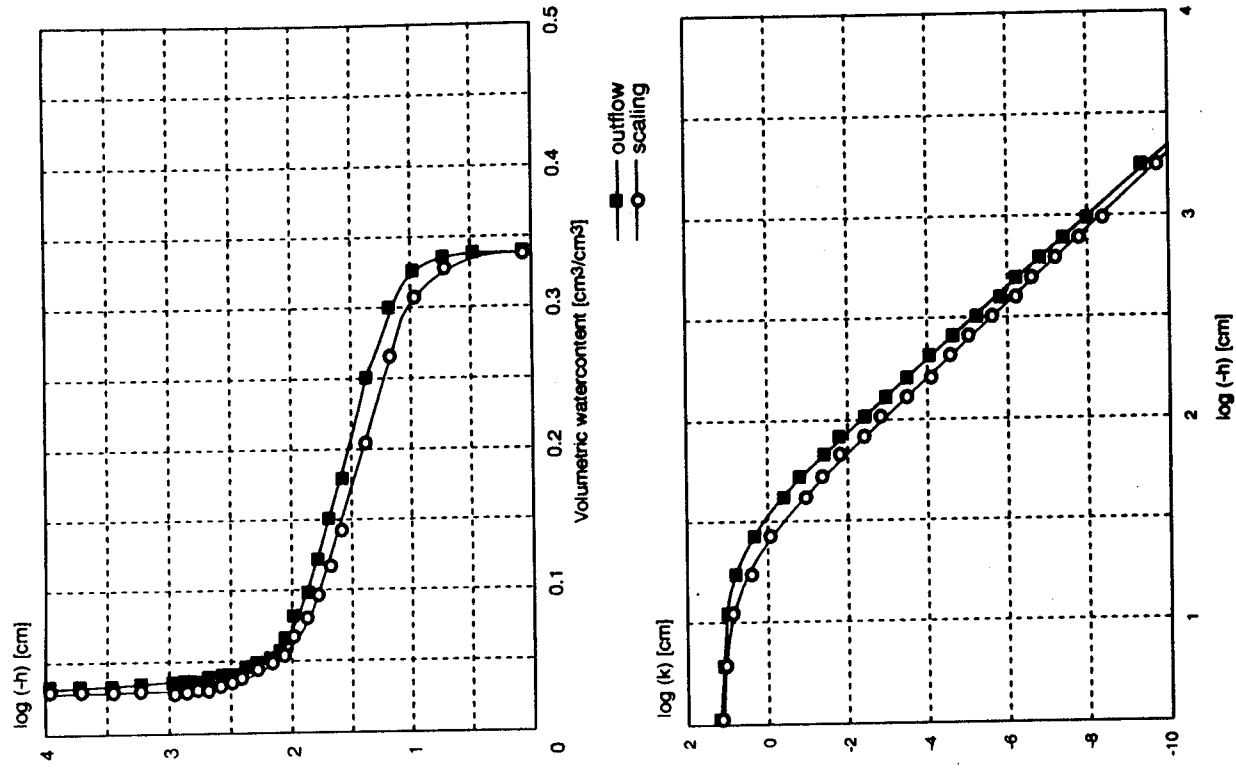


Fig 4

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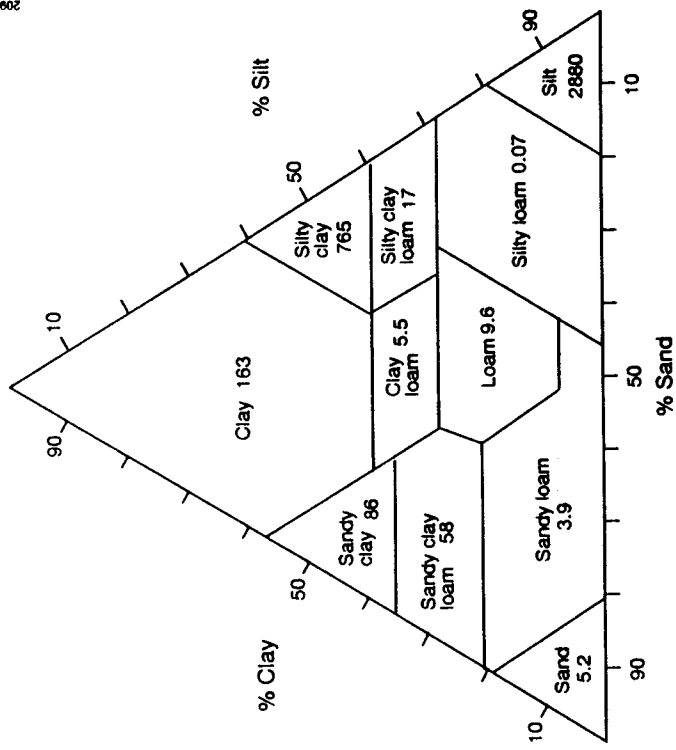


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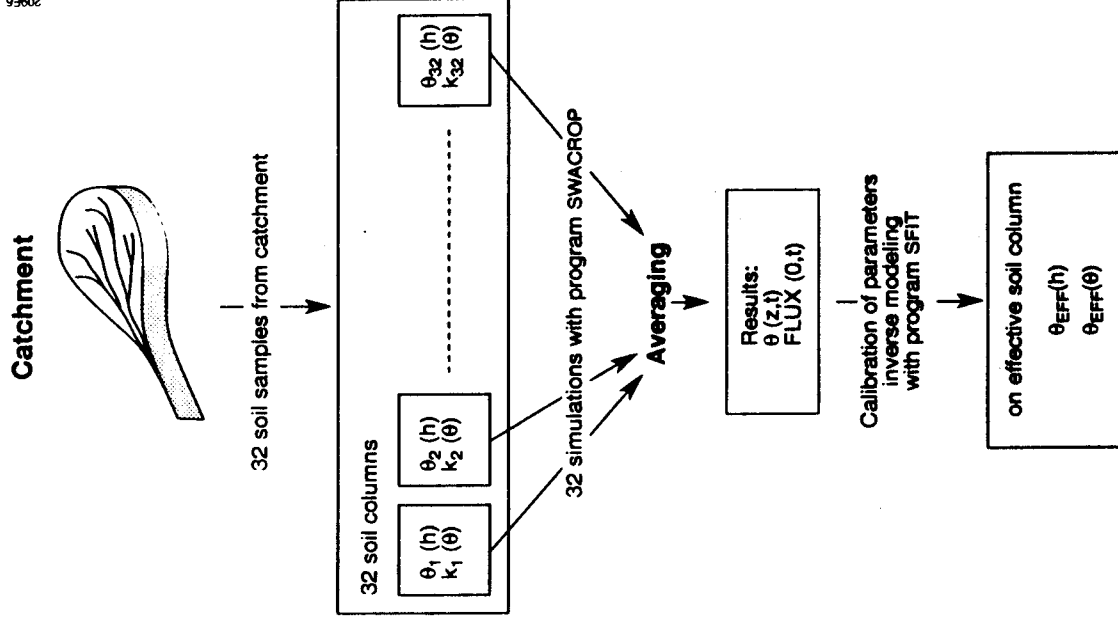


Fig 6

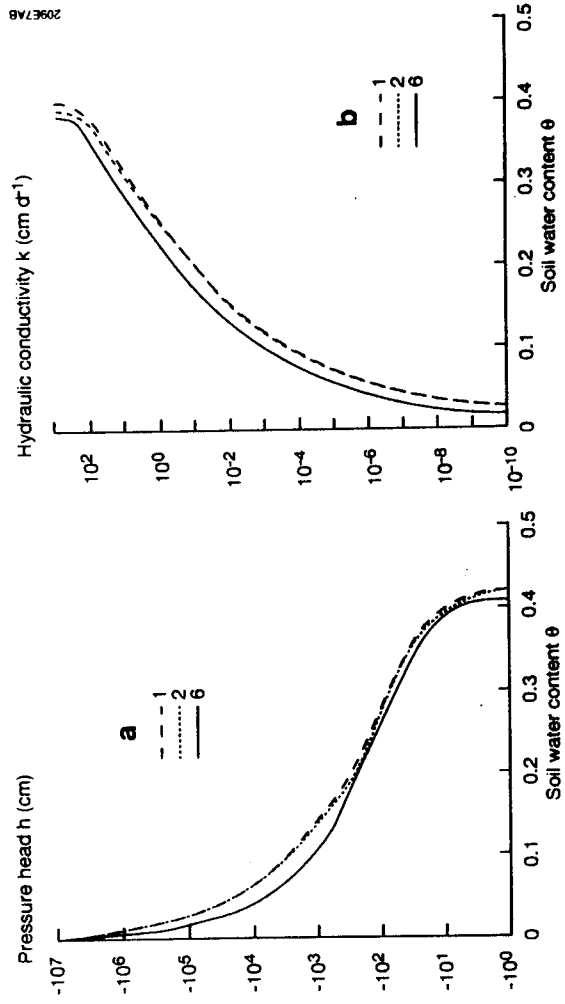


Fig 7

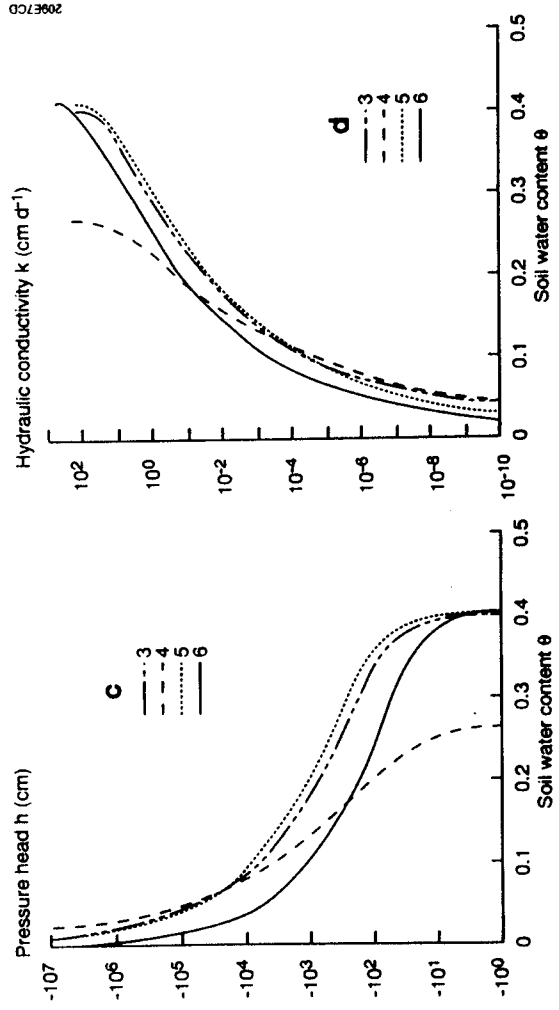
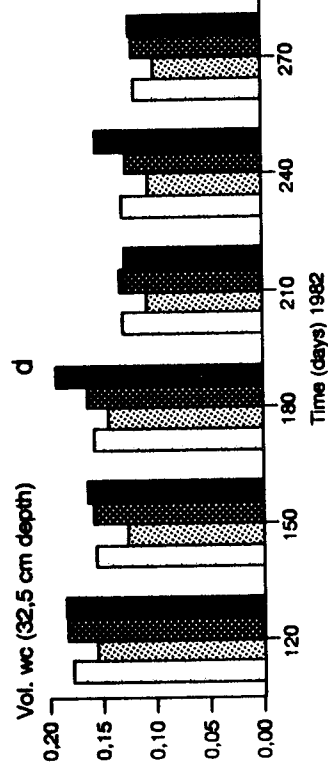
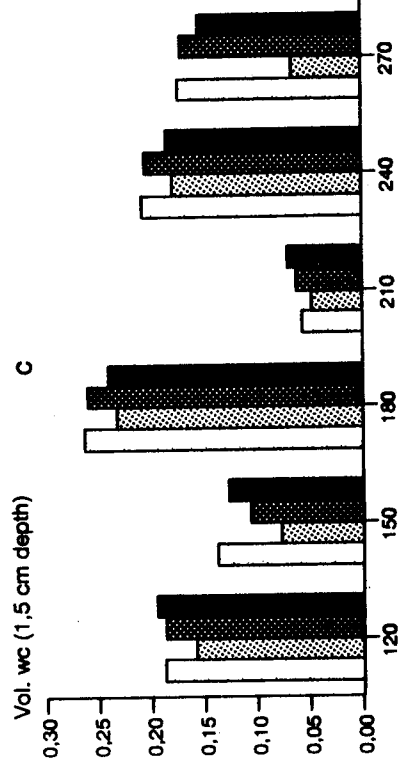
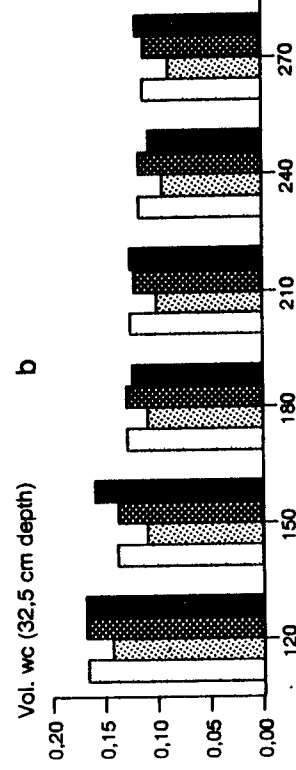
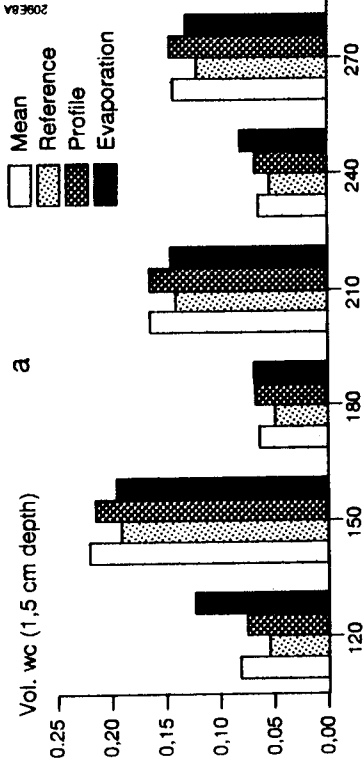


Fig 7 c, d

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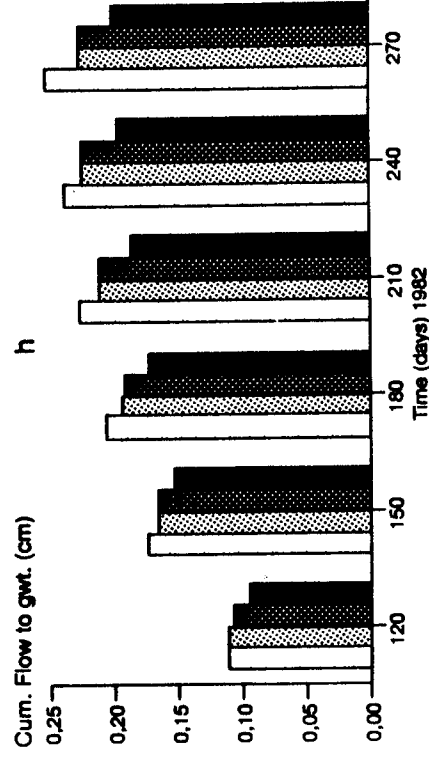
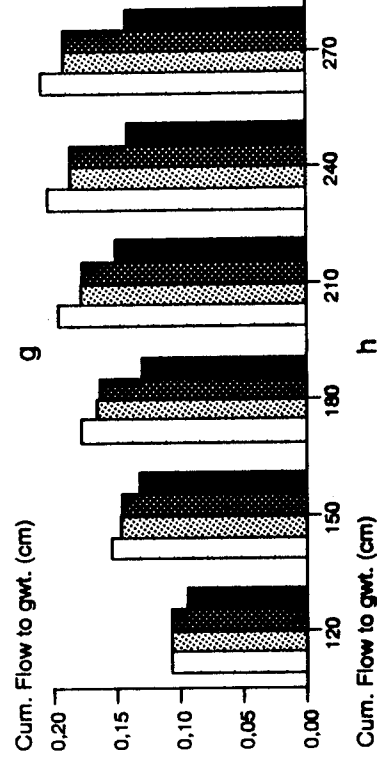
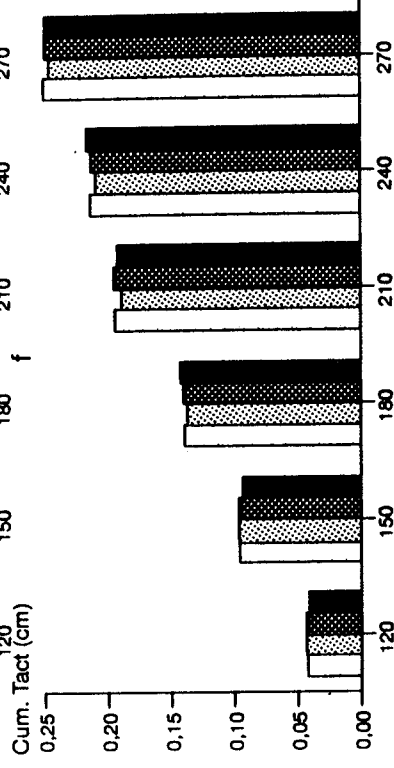
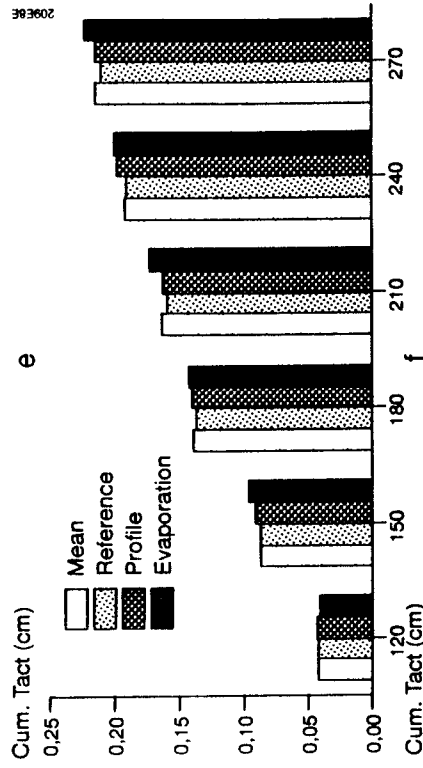


Fig 2a

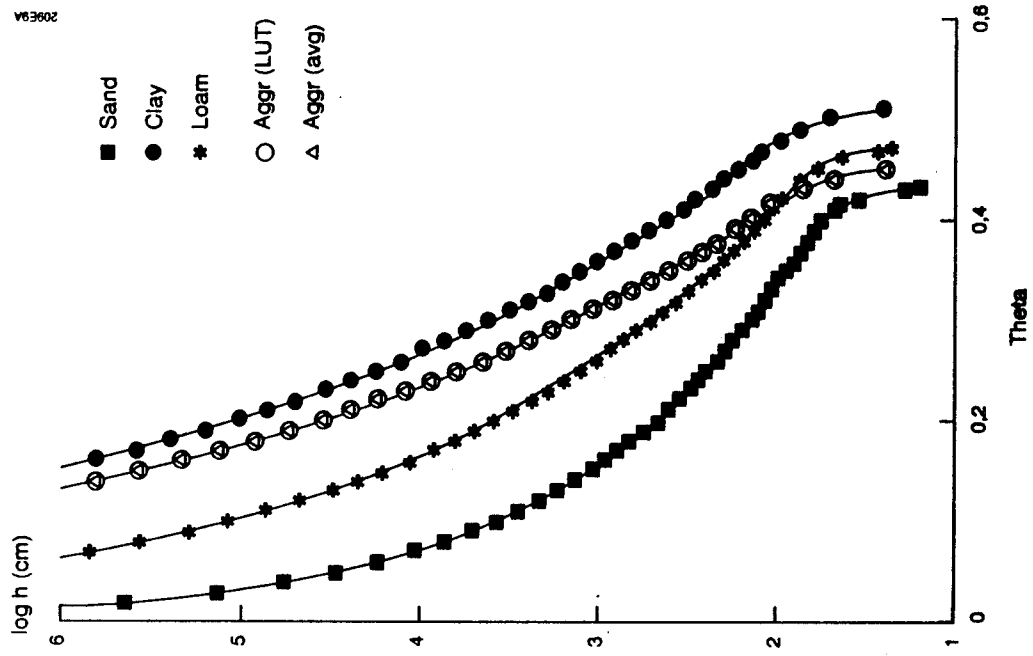


Fig 9a

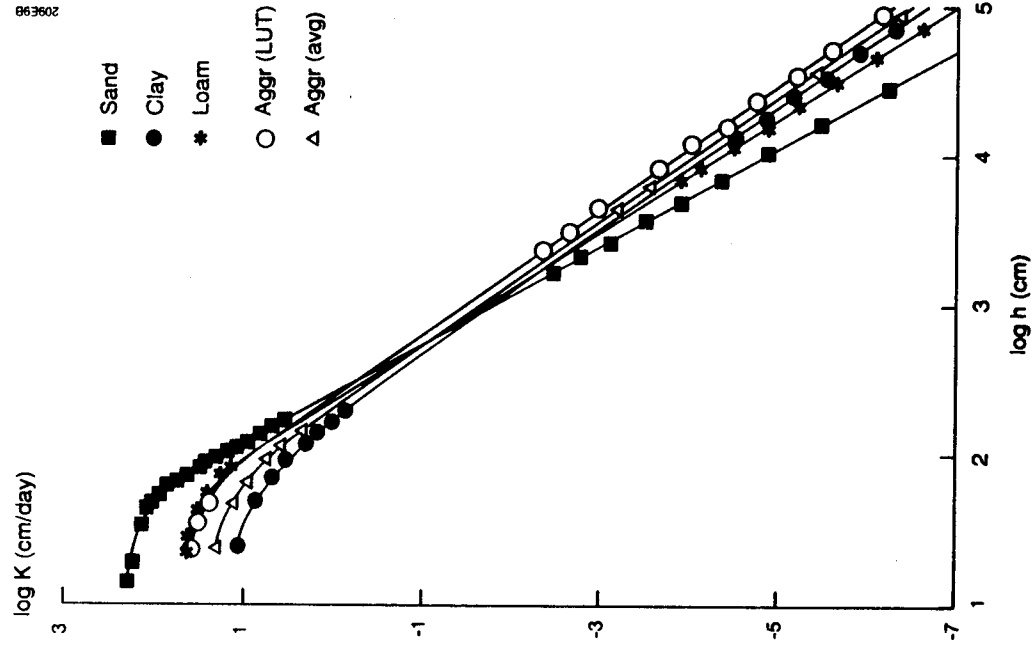


Fig 9a

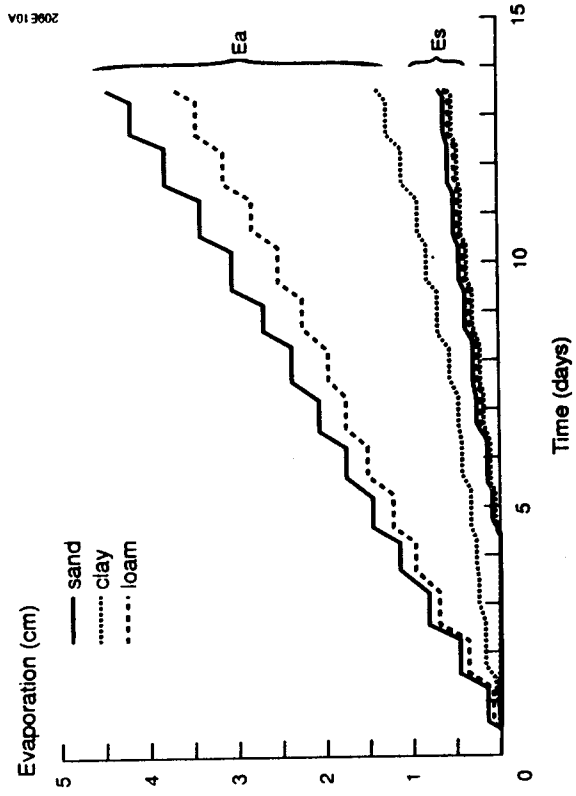


Fig 10a

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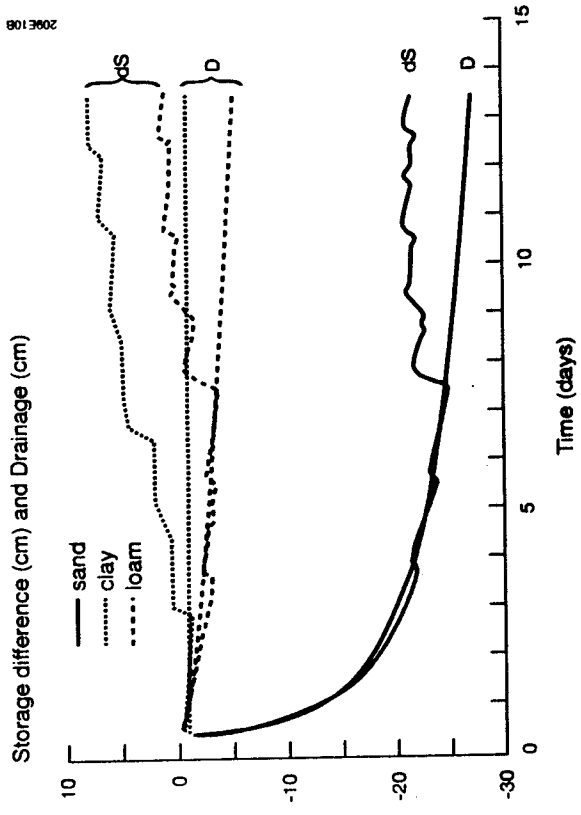


Fig 1.6

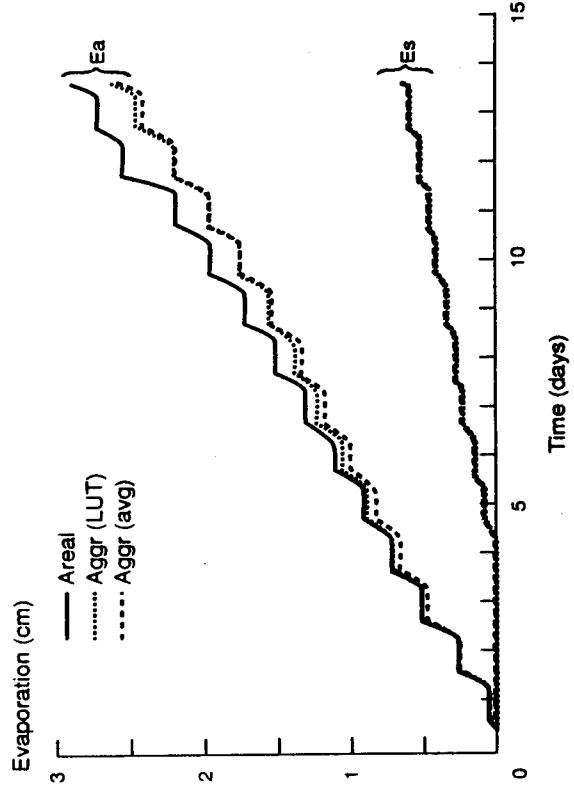


Fig 10c

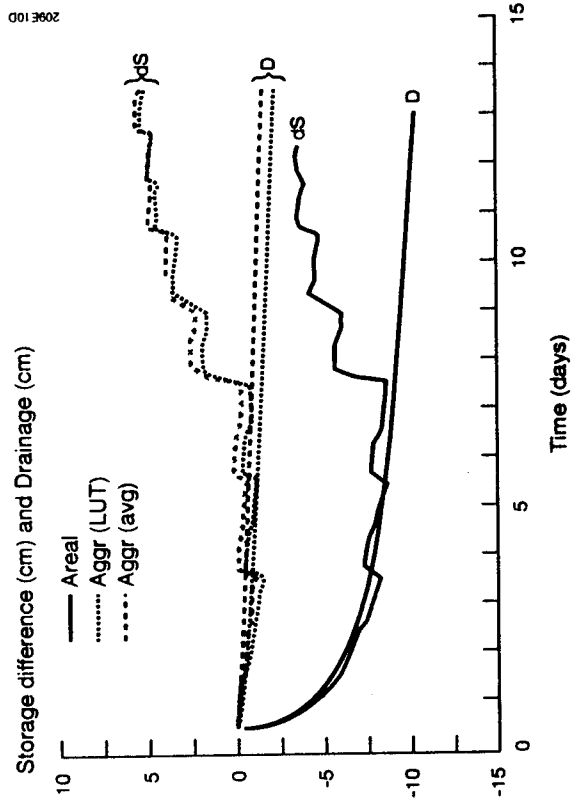


Fig 10d

THE SCALING CHARACTERISTICS OF REMOTELY SENSED VARIABLES FOR SPARSELY-VEGETATED HETEROGENEOUS LANDSCAPES

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ABSTRACT

With increasing interest in airborne and satellite-based sensors for mapping regional and global energy balance, there is a need to determine the uncertainty involved in aggregating remotely-sensed variables [surface temperature (T_s) and reflectance (ρ)] and surface energy fluxes [sensible (H) and latent (λE) heat flux] over large areas. This uncertainty is directly related to two factors: 1) the non-linearity of the relation between the sensor signal and T_s , ρ , H or λE ; and 2) the heterogeneity of the site. In this study, we compiled several remotely-sensed data sets acquired at different locations within a semi-arid rangeland in Arizona, at a variety of spatial and temporal resolutions. These data sets provided the range of data heterogeneities necessary for an extensive analysis of data aggregation. The general technique to evaluate uncertainty was to compare remotely-sensed variables and energy balance components calculated in two ways: first, calculated at the pixel resolution and averaged to the coarser resolution; and second, calculated directly at the

coarse resolution by aggregating the fine-resolution data to the coarse scale. Results showed that the error in the aggregation of T_k and ρ was negligible for a wide range of conditions. However, the error in aggregation of H and λE was highly influenced by the heterogeneity of the site. Errors in H larger than 50% were possible under certain conditions. The conditions associated with the largest aggregation errors in H were:

- sites which are composed of a mix of stable and unstable conditions;
- sites which have considerable variations in aerodynamic roughness, especially for highly unstable conditions where the difference between surface and air temperature is large; and
- sites which are characterized by patch vegetation, where the pixel resolution is less than or nearly-equal to the diameter of the vegetation "element" (in most cases, the diameter of the dominant vegetation type or vegetation patch).

Thus, knowledge of the surface heterogeneity is essential for minimizing error in aggregation of H and λE . Two schemes are presented for quantifying surface heterogeneity as a first step in data aggregation. These results emphasized the need for caution in aggregation of energy balance components over heterogeneous landscapes with sparse or mixed vegetation types.

INTRODUCTION

Remote sensing using airborne and satellite-based sensors is useful for estimating surface energy fluxes such as sensible (H) and latent (λE) heat flux over large areas. There are many algorithms and simulation models designed for this purpose. Most of these approaches utilize remote measurements of surface temperature and reflectance to compute the components of the energy balance equation [H , net radiation (R_n) and soil heat flux (G)] and estimate λE as a residual of the energy balance (see background section). Nearly all approaches require some supplementary meteorological and/or surface information.

The extent of surface information required draws a distinction between those approaches that can be applied at a local scale and those limited to regional application (Moran and Jackson, 1991). Most local scale methods rely on site-specific measurements of aerodynamic and atmospheric conditions and apply only to an area over which the ground-based measurements can be extrapolated. Regional scale methods generally rely on meteorological data from existing weather stations, extrapolated in space and interpolated in time to correspond to the location and moment of the remotely-sensed measurement.

There is interest in aggregating remotely sensed variables (and surface energy fluxes derived from these variables) from local to regional scales. This interest is due to the desire to 1) apply local-scale approaches with regional-scale data and 2) validate results of regional-scale methods by comparison with results of local-scale methods. These goals require investigation of methods for both aggregation of remotely-sensed variables

(particularly, radiometric temperature, T_r) and aggregation of energy balance components (R_n , G , H and λE).

This study addresses the issues of aggregation related specifically to heterogeneous landscapes at local and regional scales. Based on several sets of spectral images with spatial resolutions ranging from 0.3 m to 120 m, we studied the effects of aggregation by computing remotely sensed variables and energy balance components in two ways. First, the variable (e.g., H) was computed from the radiance at the pixel resolution (e.g. 120 m) and these values were averaged to obtain a value of H at a coarser resolution (e.g. 1 km). Second, pixel-resolution values of radiance were averaged to the coarser scale and then a value of H was computed. The difference between these two computations of H revealed the error due to non-linearities in the relation between surface radiance and H .

The presentation is organized into sections, with the first section presenting the basic energy balance theory that is the foundation for most local- and regional-scale models and defining the relation between kinetic, radiometric and aerodynamic temperatures. The second section discusses the issues directly related to aggregation of data for heterogeneous landscapes and suggests a method for quantifying site heterogeneity. The third section describes the study site, the sensor characteristics, and the methods for acquisition of data for this research. Finally, aggregation results are presented for a variety of homogeneous and heterogeneous sites at various scales.

BACKGROUND

Energy Balance Theory

In general, estimation of energy flux using remotely-sensed data is based on the solution of the one-dimensional surface energy balance equation, expressed as

$$R_n = G + H + \lambda E, \quad [1]$$

where R_n , G , H and λE are in units of $W\ m^{-2}$, and G , H and λE are positive when directed away from the surface. λE is a function of evaporation rate, E ($kg\ s^{-1}\ m^{-2}$), and heat of vaporization, L ($J\ kg^{-1}$)), but is typically found as a residual in Eq. [1].

R_n can also be defined as the sum of the incoming and outgoing radiant flux densities, i.e.,

$$R_n = (1-\alpha)R_{S\downarrow} + R_{L\downarrow} - R_{L\uparrow}, \quad [2]$$

where α is the hemispherical albedo, the subscripts S and L signify shortwave radiation (0.15 to $\approx 4\ \mu m$) and longwave radiation ($>4\ \mu m$) and the arrows indicate the flux direction ($\downarrow$ =incoming, $\uparrow$ =outgoing). R_n is generally estimated in one of two ways: 1) it can be measured using net radiometers and extrapolated over the region of interest; or 2) it can be

computed using remotely-sensed measurements to estimate α and R_{LH} , and ground-based instrumentation to measure the incoming terms (Jackson, 1984; Brest and Goward, 1987).

The G term is dependent upon the gradient of temperature, T ($^{\circ}\text{K}$), with soil depth, z (m),

$$G = -K(dT/dz), \quad [3]$$

where K is thermal conductivity ($\text{W m}^{-1} \text{ } ^{\circ}\text{K}^{-1}$). Results from empirical studies have shown that the daytime ratio of G/R_n is related to, among other factors, the amount of vegetation present (de Bruin and Holtslag, 1982). Thus, an approximation of G can be achieved by assuming that it is a fraction of R_n , dependent upon spectral estimates of surface vegetation cover (Moran et al. 1990; Kustas and Daughtry, 1990; Clothier et al. 1986). In some cases, such as for full-cover vegetation, G is assumed to be negligible and eliminated from the solution of Eq. [1].

The sensible heat flux density, H , is commonly expressed as a function of the difference between the aerodynamic temperature (T_a) and air temperature (T_o),

$$H = C_v(T_o - T_a)/r_a, \quad [4]$$

where C_v the volumetric heat capacity of air ($\text{J } ^{\circ}\text{C}^{-1} \text{ m}^{-3}$), and r_a is the aerodynamic resistance to heat transport (s m^{-1}). This resistance for neutral conditions (where $T_o = T_a$) is

$$r_a = \{\ln[(z-d)/z_o]\}^2/k^2U,$$

[5]

where z_o and d are roughness length and zero-plane displacement (m), respectively, U is wind speed (m s^{-1}), z is the height (m) above the surface where u is measured, and k ($=0.41$) is von Karman's constant (Brutsaert, 1982). In most cases, adjustments are made to Eq. [5] to account for non-neutral conditions ($T_o \neq T_a$) (e.g., Marht and Ek, 1984; Brutsaert, 1982; Lang et al., 1983).

Temperature Definitions

It is important at this point in the discussion to make a distinction between T_o , kinetic temperature (T_k) and radiometric temperature (T_r). Aerodynamic temperature is recognized to be the temperature at the virtual source/sink height for sensible heat exchange. The kinetic (absolute) temperature is a measure of the average kinetic energy of atomic and molecular units in motion within bodies above absolute zero. The radiometric (apparent) temperature is an estimate of the kinetic temperature, generally evaluated from a measurement of radiation emitted from the surface using an infrared sensor (Jackson, 1988).

The emittance (M_r , W m^{-2}) measured by a radiometer close to a surface (i.e., neglecting atmospheric effects in measurements acquired with an aircraft or satellite-based sensor) in a finite spectral band is related to the emittance of the surface (M_k , W m^{-2}) with the expression

$$M_r = \epsilon M_k + (1-\epsilon)M_s \quad [6]$$

where ϵ is the emissivity of the surface in the finite spectral band (assumed constant over the band) and M_s (W m^{-2}) is the incoming atmospheric emittance in the spectral band. These emittances are theoretically quantified by integrating the product of the sensor response function and the Planck function over the finite spectral band of the sensor. By integrating the Planck function over the wavelength interval 0 to ∞ , one obtains the Stefan-Boltzman law ($M = \epsilon \sigma T^4$, where σ is the Stefan-Boltzman constant: $5.674 \times 10^{-8} \text{ W m}^{-2} \text{ }^\circ\text{K}^{-4}$). For simplicity here, the Stefan-Boltzman expression is substituted for each term in Equation 6 to obtain the expression

$$\sigma T_r^4 = \epsilon \sigma T_k^4 + (1-\epsilon)B^* \quad [7]$$

In the above expression, the term B^* includes the incoming sky radiation in the spectral band 0 to ∞ and a factor related to the ratio of incoming sky radiation in the finite spectral band to incoming sky radiation in the spectral band 0 to ∞ (after Fuchs and Tanner, 1966). The emissivity term is eliminated from the left hand side of the equation because the infrared sensor is calibrated with a blackbody with emissivity close to 1. Thus, the kinetic temperature is derived from the radiometric temperature by inversion of Equation [7].

Many studies show that T_k and T_j correspond well for the case of full-cover vegetation (e.g., Huband and Monteith, 1986). However, for partially-vegetated sites, T_k and

T_o have been found to differ by as much as 5°C (Choudhury et al., 1986). From here on, T_o will refer to the temperature obtained by inverting Eq. [4], T_r will refer to the temperature measured with an infrared sensor (as described above), and T_k will refer to T_r with a correction for ϵ as described by Eq. [7]. ϵ is the ratio of emittance of a given surface to the emittance of an ideal blackbody at the same wavelength and temperature.

ISSUES RELATED TO AGGREGATION

Aggregation is simple when all relations between surface radiance and surface reflectance and temperature are linear and when relations between surface reflectance and temperature and surface energy fluxes are linear. Aggregation of remotely-sensed variables such as α and T_k becomes more complicated when their relations with radiance are non-linear due to emissivity, sensor optics and/or atmospheric conditions. The errors associated with aggregation of surface fluxes (R_n , G , H and λE) estimated with remotely-sensed variables is a function of the error associated with aggregation of remotely-sensed data and the non-linear relations between remotely-sensed inputs to the computation of surface fluxes.

This study will focus on the issues associated with aggregation of T_k and H . For the following reasons, aggregation of α , R_n , G and λE will not be addressed here:

- α : There is a great deal of published evidence that the relation between radiance and reflectance (and α) is linear for most sensors and most atmospheric conditions (Hall

et al., 1992; Humes and Sorooshian, 1994; Holm et al., 1989; Moran et al. 1994b). We obtained similar results for the surface and atmospheric conditions in this study, so presentation of such results was considered redundant.

- R_n : It is apparent from Eq. [2] that the shortwave component of R_n is a linear function of the surface albedo, which has a linear relation with radiance. The longwave components of R_n ($R_{L,1}$ and $R_{L,2}$) are related to air and surface temperatures, respectively, using the well-known Stefan-Boltzman equation (Brutsaert, 1975a). Thus, the magnitude of error in $R_{L,1}$ would be directly related to that associated with aggregation of T_k . Since aggregation of T_k will be addressed in this study, it was deemed unnecessary to repeat the analysis for $R_{L,1}$. Actually, the variables that are associated with most of the error in aggregation of R_n are the incoming terms ($R_{S,1}$ and $R_{L,1}$) which are generally estimated with ground-based instrumentation and are thus unrelated to remotely-sensed data.

- G : There is evidence that the relation of G/R_n is linearly or near-linearly related to a spectral vegetation index computed from the ratio of the near-IR and red reflectances (Kustas and Daughtry, 1990; Clothier et al. 1986; Moran et al., 1990). Again, since previously-published studies reported a linear relation between reflectance and radiance, there is little chance for error in the aggregation of values of G from local to regional scales.

- λE : Since λE is computed as a residual in the energy balance equation (Eq. [1]), the sum of the errors associated with aggregation of R_n , G and H would equal the error in λE . And since we are assuming negligible error in aggregation of R_n and G , the results of our study of aggregation of H are directly applicable to estimates of λE .

In the following subsections, we will address the issues related to aggregation of T_k and H and the relation of aggregation error with surface heterogeneity.

Aggregation of Temperature Data

The relation between radiance measured by a satellite sensor and values of surface temperature is not necessarily linear. For example, the relation between T_r and spectral radiance (L) is given for Landsat Thematic Mapper (TM) as

$$T_r = K_2 / [\ln(K_1/L + 1)], \quad [8]$$

where K_2 ($^{\circ}K$) and K_1 ($mW\ cm^{-2}s^{-1}\mu m^{-1}$) were determined for Landsat4 and Landsat5 TM by Markham and Barker (1986). This equation is a close approximation of the integration of the Planck function over the spectral bands of the TM sensor. Because temperature is a slightly non-linear function of observed radiance, the aerielly-integrated temperatures

derived from radiance observed with sensors of low- and high-resolution over the same area may be slightly different.

Furthermore, the radiometers provide only an estimate of the radiometric temperature with no correction for surface emissivity. As shown in Equation [7], an emissivity less than unity decreased the apparent temperature, whereas the reflection of incoming sky radiation increased it by a different amount (Fuchs and Tanner, 1966). Thus, the data collected using ground- or satellite-based sensors must be corrected for the surface emissivity. For a series of natural surfaces within a savanna environment, Van de Griend and Owe (1993) found that ϵ could be estimated as

$$\epsilon = 1.0094 + 0.042(\ln(\text{NDVI})), \quad [9]$$

where NDVI is the Normalized Difference Vegetation Index $[(\rho_{\text{NIR}} - \rho_{\text{red}})/(\rho_{\text{NIR}} + \rho_{\text{red}})]$ and ρ_{NIR} and ρ_{red} are the near-IR and red reflectances, relatively. The NDVI was derived to be sensitive to changes in vegetation cover (Richardson and Wiegand, 1977); thus, it appears that the relation between ϵ and vegetation cover is logarithmic.

Aggregation of Sensible Heat Flux

For relatively homogenous landscapes, Hall et al. (1992, Section 3.2.1) found that the relation between remotely sensed TM data and radiance was sufficiently linear to allow H

values derived with TM-resolution (30 m or 120 m) remotely-sensed data to be averaged to produce low-resolution (1 km) values with little error. However, based on a theoretical evaluation of the eddy diffusion formulation of H (Eq. 4), they predicted a breakdown in scale invariance for heterogeneous landscapes (their section 2.4.1). Scale invariance may not hold for heterogeneous landscapes (such as semi-arid rangeland) for three reasons. First, substantial variations in vegetation type, height and cover associated with heterogeneous landscapes result in large variations in resistance to heat transfer (Monteith 1973; 1981). There is also evidence that resistance is dependent upon the magnitude of the difference between the soil and foliage temperature (Kustas et al., 1989). In semiarid rangelands, the difference between soil and vegetation temperature can be as large as 40°C (Humes et al., 1994b). Thus, one would expect a greater error in the averaged values of H and λE for a heterogeneous site than for a homogeneous landscape.

Second, heterogeneous landscapes would likely include surfaces with both stable ($T_o < T_a$) and unstable ($T_o > T_a$) conditions. Thus, the neutral equation for resistance must be corrected for effects of stability. In most formulations, the stability correction results in a non-linear relation between T_o and r_a and, consequently, would result in a non-linear relation between fluxes aggregated from fine-resolution data and those computed with coarse-resolution data.

Third, there is a non-linear relation between T_k and T_o over a range of vegetation densities. That is, the difference between T_o and T_k is small for fully-vegetated surfaces (or sites with little difference between soil and vegetation temperatures) and increases rapidly

with increasing heterogeneity of the surface (when the difference between soil and vegetation temperatures is large). For a surface only partially covered by vegetation (where T_k is a composite of the soil and vegetation temperatures), Kustas et al. (1990) reported that the resistance to heat transfer was significantly influenced by rapid changes in the soil surface temperature. Thus, observed values of H several meters above the surface changed relatively little compared to changes in $T_k - T_a$ (Kustas et al., 1989). Consequently, for partially-vegetated surfaces, it is necessary to incorporate an additional or "excess" resistance to sensible heat transfer.

Several studies have addressed the computation of an excess resistance term based on the fact that momentum transfer (based on pressure and viscous forces) is more efficient than transfer of scalar quantities, such as H , which are transferred by viscous forces only (Brutsaert, 1975b). This results in additional resistance to heat transfer that can be expressed in the resistance equation by including different roughness lengths for momentum (subscript "m") and heat (subscript "h"), where

$$r_a' = \{ [\ln((z-d)/z_{om}) + \ln(z_{om}/z_{oh}) - \psi_h] [\ln((z-d)/z_{om}) - \psi_m] \} / k^2 U, \quad [10]$$

and z_{om} and z_{oh} are the roughness length for momentum and scalar roughness for heat, respectively, and ψ_h and ψ_m are the stability corrections for heat and momentum (summarized by Beljaars and Holtslag, 1991). A kB^{-1} factor [$\text{defined } kB^{-1} = \ln(z_{om}/z_{oh})$] was

included in the resistance computation, resulting in a non-linear relation between $(T_k - T_a)$ and H . Thus, it is possible to rewrite Eq. [4] as

$$H = C_v(T_k - T_a)/r_a'. \quad [11]$$

In application, Kustas et al. (1989) suggested that kB^{-1} could be computed as a function of wind speed (U) and $(T_k - T_a)$ and proposed that

$$kB^{-1} = s_{kb} U(T_k - T_a), \quad [12]$$

where s_{kb} is site-specific and was computed as 0.17 for Owens Valley rangeland and 0.13 for Walnut Gulch rangeland (Kustas et al., 1994b).

There is substantial evidence supporting the importance of accounting for the difference between T_o and T_k in Eq. [4] (e.g., Choudhury et al., 1986). Though inclusion of the kB^{-1} factor in the resistance term is only one of several approaches that have been proposed to account for the difference between T_o and T_k (others include Chehbouni et al., 1994; Vidal and Perrier, 1990; Norman et al. 1994; Shuttleworth and Wallace, 1985; Kustas, 1990; Moran et al. 1993; Prevot et al., 1993), it was adopted for use in this study due to extensive studies that have proven its application for semi-arid rangeland (Moran et al., 1994c; Humes et al., 1994a; Kustas et al., 1994b, 1994c; Stewart et al., 1994).

Relation of Surface Heterogeneity to Aggregation

The issues identified above for aggregation of T_r and H are inextricably linked to the heterogeneity of the site. For example, the sensor calibration presented in Eq. [8] is nearly-linear for a large range of T_r values, resulting in little error in aggregation for homogeneous sites with a low range of T_r values. The same is true for the error associated with emissivity, where the logarithmic relation between ϵ and NDVI will produce little scaling error for uniformly-vegetated sites with a low range of NDVI values. And the scaling errors associated with the stability corrections in Eq. [10] will produce the greatest scaling errors for heterogeneous sites with an equal distribution of stable and unstable conditions. Thus, it is important to quantify the heterogeneity of the site before attempting aggregation of the remotely-sensed variables or energy balance components.

The importance of quantifying heterogeneity is matched by its difficulty. The variability of remotely-sensed variables and energy balance components for a site is influenced by surface-related factors such as vegetation type/cover and soil moisture, and other factors such as time of day, time of year and spatial resolution. This complexity leads to some misconceptions about site heterogeneity. For example, at one time of day at one site, the variability in T_r for a sparsely-vegetated rangeland site would probably be less during the "dry" season than during the "wet" season when vegetation and soil moisture are highly variable. However, an early-morning measurement during the wet season would have less variability than a mid-day measurement during the dry season (due to differential

heating of vegetated and bare surfaces). Furthermore, a fine-resolution image during the dry season would likely show more variability than a coarse-resolution image during the monsoon season. Thus, it is erroneous to make the assumption that all sites will be heterogeneous during the wet season and homogeneous during the dry season.

In this study, we used to two methods to quantify the heterogeneity of a site. The first was simply a comparison of the shapes, magnitudes and ranges of histograms formed from image data. The second method utilized a scattergram of the NDVI versus surface-air temperature ($T_k - T_a$) from the spectral image. The significance of the latter requires some theoretical background which will be provided here, though readers are encouraged to refer to previously-published works (Price, 1990; Gillies and Carlson, 1994; Nemani and Running, 1989; Moran et al., 1994a).

A scattergram of the spectral vegetation index (e.g. NDVI or soil-adjusted vegetation index (SAVI), Huete, 1988) vs. $T_k - T_a$ generally results in a trapezoidal shape that is indicative of the variation in vegetation cover and evapotranspiration (e.g., Figure 1). Carlson et al. (1990) associated the scatter with differences in root-zone moisture availability (5-100 cm), soil surface moisture (0-2 cm) and fraction of vegetation cover. The shape of the scatter is thus related to the heterogeneity of the site. Boundaries can be drawn encompassing the maximum-possible variability in the vegetation index and $T_k - T_a$, allowing determination of the absolute heterogeneity of the site defined by meteorological conditions. Moran et al. (1994a) proposed a theoretical basis for computing the maximum and minimum

possible surface temperatures for a range of vegetation cover from 0 to 100% using a variation of the Penman-Monteith equation,

$$(T_k - T_a) = [r_a' (R_n - G) / C_v] \{ [\gamma (1 + r_d / r_a') / \{ \Delta + \gamma (1 + r_d / r_a') \}] - [VPD / \{ \Delta + \gamma (1 + r_d / r_a') \}] \}, \quad [13]$$

where r_c the canopy resistance ($s\ m^{-1}$) to vapor transport, γ the psychrometric constant ($kPa\ ^\circ C^{-1}$), Δ the slope of the saturated vapor pressure-temperature relation ($kPa\ ^\circ C^{-1}$), and VPD the vapor pressure deficit of the air (kPa). This work resulted in the derivation of a Vegetation-Index/Temperature (VIT) Trapezoid that encompassed the maximum possible variability in the vegetation index and values of $(T_k - T_a)$ for one site on one date (Figure 1).

The VIT Trapezoid theory defines $T_k - T_a$ for four extreme situations by changing the r_c term in Eq. [13]:

- 1) full-cover, well-watered vegetation, where r_c is the canopy resistance at potential evapotranspiration (assumed to be $5\ s\ m^{-1}$);
- 2) full-cover vegetation with no available water ($\lambda E = 0$), where r_c is the canopy resistance associated with nearly complete stomatal closure (assumed to be $1000\ s\ m^{-1}$);
- 3) saturated bare soil, where $r_c = 0$ (the case of a free water surface); and
- 4) dry bare soil, where $r_c = \infty$ (analogous to complete stomatal closure).

The left edge of the trapezoid was related to conditions of potential evapotranspiration and the right edge was related to conditions in which $\lambda E=0$. Thus, the location of the image data within the trapezoid provides information about the variability of H and λE , in addition to the basic information on variability of the vegetation index and T_k-T_a . The VIT trapezoid will be used in subsequent sections to illustrate the heterogeneity of the study sites.

Summary of Issues

In order to test any schemes for aggregation of T_r and H, variability in the following points must be considered:

- 1) sensor calibration;
- 2) surface emissivity;
- 3) site aerodynamic stability;
- 2) z_0 between sites; and
- 3) z_{om} and z_{oh} at each point.

Furthermore, these issues are interrelated and effects of each are dependent upon the variability of other features. Thus, it is equally important to consider the heterogeneity of the site in any aggregation scheme. Using an extensive set of remotely-sensed and meteorological data acquired in a semiarid rangeland in Arizona during the dry and wet

season, a first attempt will be made to define the sensitivity of linear aggregation schemes to these points.

DATA SETS AND METHODS OF ANALYSIS

An experiment was conducted at the Walnut Gulch Experimental Watershed (WGEW) near Tombstone, Arizona, to acquire the meteorological and remote sensing data necessary to evaluate the scaling characteristics of heterogeneous landscapes. These data were acquired as part of larger study (Monsoon'90) focusing on the general utility of remote sensing to provide a practical means for monitoring some of the important factors controlling land surface processes (Kustas et al., 1991). The experimental sites were located in an area comprising the upper 100 km² of the Walnut Gulch drainage basin, from about 1300-1800 m above mean sea level (MSL). In this region, precipitation ranges from 250-300 mm/yr, with 2/3 of the rainfall occurring during the summer "monsoon season" in July and August. For this study, data were obtained during the dry season in June while most vegetation was still dormant, and during the monsoon season in late July and early August, when the vegetation was at peak greenness and soil moisture was highly variable in time and space due to recent precipitation events (Schmugge et al., 1994).

Most of the analysis was limited to eight sites within the watershed, termed METFLUX stations (herein referred to as MF1 to MF8), containing instrumentation for measuring both general meteorological conditions and estimating the surface energy balance

(Kustas et al., 1994a). The METFLUX sites were located along two parallel transects crossing the two dominant vegetation types (brush and grasses) and the transition from one to the other (Figure 2). Two of the eight METFLUX sites contained considerably more micrometeorological instrumentation than the others: MF1 (located in the Lucky Hills subwatershed) was located in a relatively flat, brush-dominated ecosystem and MF5 (Kendall subwatershed) was in a hilly, grass-dominated subwatershed.

Some analysis was conducted for a site just southeast of WGEW that included a distinct riparian zone, characterized by lush perennial vegetation. The riparian zone presented a linear feature in high contrast with surrounding brush and grassland vegetation, resulting in very heterogeneous conditions.

Data Sets

Data from four sensors mounted on four different platforms were used for this analysis. The platforms, sensor characteristics, locations of ground targets, and dates of data acquisition are summarized in Table 1. A more detailed account of data acquisition methods and site descriptions follows.

Yoke-based radiometers: Radiometers were mounted on backpack-type devices (yokes) to be carried by operators at two sites (Lucky Hills and Kendall) to characterize by sparse shrubs and grass, respectively. At both sites, a large ground target was

delimited over which surface temperature and reflectance were measured from a height of 2 m above the ground surface, using yoke-based radiometers and a calibrated reference reflectance panel (Jackson et al., 1987). Data were acquired at 1 m increments along transects through the target, covering the entire target in less than 15 minutes (this technique was similar to that described by Slater et al., 1987). The Lucky Hills target was approximately 120 by 120 m in size, typified by relatively flat topography and primarily shrub vegetation of ≈ 0.6 m height covering 20% of the soil surface. The Kendall target was much larger, 480 by 120 m, located in a hilly, grass-dominated site, stretching from the top of one hill eastward to the top of another. Yoke-based spectral data were processed to produce reflectance and temperature for individual samples (0.3 m resolution) on days of year (DOYs) 156 and 252 (to correspond with TM data) at 10:30 am at Lucky Hills and Kendall.

Thermal Scanner: Thermal infrared data were acquired over WGEW with an Inframetrics 600 thermal imaging radiometer during the Monsoon'90 Experiment. The scanner was flown in a single engine Cessna at altitudes ranging from 0.09 km to 3.35 km above ground level (AGL) on several days from DOY 209 to DOY 222. A subset of these data were selected coinciding with clear-sky conditions and a variety of surface conditions. Thermal data were selected for Kendall and Lucky Hills on three dates at 0.09 km AGL: DOY 209 @ 10:12, DOY 209 @ 14:40 and DOY 221 @ 7:30. Thermal data acquired at higher altitudes, 0.92 km and 3.35 km AGL, were selected

for Lucky Hills on DOY 222 @ 14:11 and Kendall on DOY 209 @ 14:51, respectively. In all cases, a window of size 380 by 480 pixels was extracted surrounding the designated METFLUX site. Thus, at a flight altitude of 0.09 km AGL, the window covered an area of 76 by 96 m and at 3.35 km AGL the window covered 2.3 by 2.9 km.

NSO01: The C-130 aircraft with the NS001 sensor flew at an altitude of 2 km AGL along transects intersecting the locations of the 8 METFLUX sites. Windows of size approximately 2.3 by 2.9 km were extracted for areas surrounding each of the 8 METFLUX sites. Though the NS001 sensor provides data in 8 spectral bands, at this time only the thermal data were available. We limited our analysis to data collected during a mid-day flight on DOY 221 with cloudfree conditions.

Landsat TM: Landsat-5 TM data were acquired on DOY 156 (dry season) and 252 (post-monsoon season) covering an area 180 by 180 km, encompassing WGEW. Analysis was conducted for two adjacent areas of size 10 by 3 km within the scene. One area covered most of WGEW and another covered an area just southeast of WGEW that included very diverse terrain and some pixels with distinctive riparian vegetation (this area will be referred to as RIP to distinguish it from WGEW). Prior to any tests of aggregation, the reflected data (TM1-TM4) were averaged from 30 m resolution to 120 m resolution to provide reflected and emitted data at the same resolution.

We emphasize that these data sets were selected for analysis because they covered a wide range of conditions due to differences in spatial resolution, time of acquisition, and site characteristics. The study presented here is not meant to be an evaluation of the effects of *spatial resolution and other data-specific characteristics* on aggregation, but rather an evaluation of the effects of *surface heterogeneity* on aggregation. The differences in the spatial resolution, timing, and location of these data sets resulted in differences in the heterogeneity of the surface temperatures and reflectances that were necessary for an extensive analysis of data aggregation. Though we realize that most interest in aggregation is at the regional and global scales, the aggregation of our finest resolution data (less than plant diameter) provided the opportunity for insight into the uncertainty of aggregation at coarser scales (e.g., sites characterized by distinctive patch vegetation).

Methods of Analysis

All analyses had the same general form. Values of T_r and H were each computed in two ways:

Surface temperature: First, T_r was computed from radiance at the pixel resolution (e.g. 120 m for TM) and corrected for emissivity (using Eqs. [7] and [9], where B^* was assumed to be zero and NDVI was computed for each pixel), and these values were averaged to obtain a value of T_k at a coarser resolution (e.g. 1 km for TM). Second,

pixel-resolution values of radiance were averaged to the coarser scale (e.g. 1 km for TM) and then a value of T_r was computed and corrected for emissivity (based on an average NDVI value in Eq. [9]) to estimate T_k . All tests in which visible and near-IR data were available (yoke and TM), temperatures were corrected for emissivity using Eqs. [7] and [9]. In all other tests, we assumed that $\epsilon=1.0$ for all temperatures.

Sensible heat flux: First, H was computed from T_k at the pixel resolution (Eqs. [10] - [12]) and these values were averaged to obtain a value at the coarser resolution. Second, pixel-resolution values of T_k were averaged to the coarser scale, and then a value of H was computed using Eqs. [10] - [12] based on average values of T_k . All tests were conducted using resistance equations that included the kB-1 factor, using $s_{kB}=0.17$. In all cases, corrections were made for stability conditions; that is, different equations were used for stable and unstable conditions, according to Beljaars and Holtslag (1991). Meteorologic inputs (z , U and T_a) were taken from a local station for the time of the spectral measurements, and values of z_o and d for these sites were computed by Kustas et al. (1994a).

For all data sets except TM, the data were aggregated from pixel-resolution to a single point covering the entire extracted window. The TM data were aggregated from 120 m to 1 km resolution over the window covering a 10 by 3 km area. Thus, the aggregation of TM

data resulted in 30 (that is, 10x3) pixels, and the aggregation for all other data sets resulted in a single value for each.

RESULTS - SITE VARIABILITY

To investigate the variability of the vegetation index and T_r - T_a for the finest (yoke) and coarsest (TM) resolutions, the VIT Trapezoid was computed for WGEW based on on-site meteorological data for DOY 156 and 252. As described previously, the left edge of the trapezoid is related to conditions of potential evapotranspiration and the right edge is related to conditions in which $\lambda E=0$. The yoke-based data illustrated the variability at sub-element scale (0.3 m resolution < plant diameter) and the TM data illustrate variability of mixed pixels (120 m resolution >> plant diameter). For DOY 156 (dry season), the yoke and TM data were clustered in the lower right corner of the VIT Trapezoid, indicating low vegetation cover and dry conditions (Figure 3a). The variability of the yoke-based NDVI and T_k - T_a data was much larger than that of the TM data, even though the area of coverage was much smaller. For the TM data, the variability of the WGEW site was lower than that of the RIP site which encompassed a distinct riparian zone.

For DOY 252, the yoke and TM data were shifted toward the left edge of the trapezoid, indicating higher λE values associated with the wet season (Figure 3b). The general variability of NDVI and T_k - T_a was much greater than for DOY 156 at both resolutions. The variability of the yoke data at Lucky Hills (MF1) was greater than the same

data for Kendall (MF5). Similar to results for DOY 156, the variability of the yoke data was greater than that of the TM data, and the variability of the TM data for the RIP site was greater than that for WGEW.

When only thermal data were available, we were unable to use the NDVI vs. $T_k - T_a$ scattergram to evaluate site heterogeneity. Instead, we evaluated the relative site heterogeneity by comparing the histograms of $T_r - T_a$ from images acquired at different times of day and locations. Since thermal scanner images were acquired at several locations, altitudes, times of day, and times of season, it was possible to investigate the influence of these factors on the variability of $T_r - T_a$:

- Time of Day: The range of $T_r - T_a$ values increased with time of day at both Lucky Hills and Kendall (Figure 4a and Table 2). This was due to the differential thermal inertia and evapotranspiration of vegetation and soil, and it generally results in greater absolute variability near noon. However, since one of the issues to be addressed in aggregation of H is the non-linear relation between $T_k - T_a$ and atmospheric stability, we must also consider the mix of unstable and stable conditions. The early morning measurements tend to have a greater mix of stable and unstable conditions; later readings tended toward uniformly unstable conditions.
- Time of year: Images acquired on DOY 209 (early-monsoon) and DOY 221 (mid-monsoon) for the same site and time of day were used to investigate the influence of

seasonal vegetation differences (Figure 4b and Table 2). Though the range of $T_r - T_a$ values decreased slightly with the increase in soil moisture and vegetation cover associated with the monsoon season, the most distinctive feature of these data was the shift of the histogram peak by nearly 5 °C, associated with more actively transpiring vegetation component on DOY 221.

- Vegetation type/cover: The variability associated with vegetation type and cover was illustrated by comparison of $T_r - T_a$ values for the Kendall and Lucky Hills sites on the same day and time, with the same spatial resolution (Figure 4c and Table 2). Kendall was characterized by a uniform distribution of grass; whereas, Lucky Hills was characterized by scattered shrubs and bare soil. Consequently, the histogram of data from Kendall formed a unimodal distribution skewed toward higher temperatures, and the histogram of Lucky Hills' data tended toward a bi-modal shape with peaks associated with the shrubs and bare soil (Humes et al., 1994b).

- Spatial resolution: Thermal data acquired near the same time-of-day at two spatial resolutions (0.2 m and 6.0 m) with a field of view centered on Kendall were compared (Figure 4d and Table 2). As expected, the coarser resolution data had less variability than the finer resolution data; the range of $T_r - T_a$ values decreased from 27 °C (at 0.2 m resolution) to 13 °C (at 6.0 m resolution). This was the case even though the coarser-resolution image covered a ground surface area that was larger and

more variable than the smaller area covered by the finer-resolution image. Thus, it is apparent that at these scales the highest variability in $T_r - T_a$ is associated with the finer resolutions.

It should be noted that this analysis was conducted without correction of the thermal data for atmospheric attenuation. As precipitable water increases, ground temperatures detected by airborne and satellite sensors tend to converge (Kiang, 1982). Consequently, the overall effect of the atmosphere is to reduce the thermal image contrast (Byrnes and Schott, 1986). This would affect the results presented in Figures 3 and 4, where data acquired on different days, at different sites, or at different altitudes were directly compared. This effect could partially explain the decrease in variability of the satellite data relative to that of the yoke-based data (Figure 3) and the decrease in variability with sensor resolution (Figure 4d). The atmospheric influence would have less effect on the following analysis in which aggregation techniques are applied to single images and the within-image variation is presented.

RESULTS - AGGREGATION

In previous sections, we identified several sources of non-linearity in the relation between radiance and T_r (due to sensor calibration), between T_r and T_k (due to emissivity), and between $T_k - T_a$ and H (due to stability corrections of r and variability in surface

roughness, z_0). Here, we used data from the yoke-based radiometers, aircraft-based thermal scanner and NS001, and satellite-based TM to investigate the effects of these non-linearities on aggregation of T_k and H . As described previously, the general technique to evaluate the error was to compare T_k and H values calculated in two ways: first, calculated at the pixel resolution and averaged to the coarser resolution; and second, calculated directly at the coarse resolution by aggregating the fine-resolution data to the coarse scale and then computing T_k or H .

Variability in Aggregated Values of T_k due to Sensor Calibration and Surface Emissivity

The relation between T_r and radiance measured by the TM sensor is nearly linear over a large range of T_r values and resulted in aggregation errors less than 0.05°C for temperature ranges of 30°C for prairie vegetation in Kansas (Humes and Sorooshian, 1994) during FIFE, the First ISLSCP Field Experiment (Sellers et al., 1988). Similar results were found here for the TM data acquired for WGEW on DOYs 156 and 252. Since this error again proved to be both small and quantifiable (using Eq. [8]), it will not receive further discussion.

The error in aggregation of T_k associated with the logarithmic relation between emissivity and NDVI (Eq. [9]) was investigated using the TM and yoke-based data sets. The error in aggregation associated with the emissivity correction (and including the non-linearity in the sensor calibration) was minor for the TM-resolution data over the range of surface temperatures associated with two sites and two dates (Figure 5). The average difference in

T_k values using the two aggregation schemes was less than 0.1°C . The error in aggregation associated with the emissivity correction was also minor for the yoke-based data at a much finer resolution (0.3 m) than TM with a larger range of T_r values. In this case, the differences in T_r using the two aggregation schemes were less than 0.04°C .

These results lead to the conclusion that error associated with aggregation of T_k due to emissivity differences was negligible for both fine and coarse resolutions for relatively homogeneous sites. The error increased slightly (though it was still less than 0.1°C) for heterogeneous sites, such as RIP on DOY 252.

Variability in Aggregated Values of H due to the Stability Correction of r_a

In the following tests, we assumed that z_o was constant at both the pre- and post-aggregation scale. In this way, we could investigate the variability in H associated with variability in the stability correction of r_a' and variability in kB^{-1} . By comparing results at different sites and different resolutions, we tested the magnitude of error associated with magnitude of z_o , stability correction, and emissivity.

At fine resolution (0.3 m for yoke-based and low-altitude thermal scanner data), error in aggregation of H associated with the stability correction of r_a could be large, depending on surface conditions. For the yoke-based data, errors as large as 12 W m^{-2} were found for highly variable sites such as Lucky Hills during the wet season (Figure 6). Error was smaller for sites with less variability, such as the Kendall grassland during both the wet and dry

seasons. It is notable that the site conditions during data acquisition of yoke-based data (@ 10:30am) were dominated by unstable conditions.

In contrast, the low-altitude thermal scanner data (with footprint of 0.3 m) was acquired during morning, mid-day and afternoon, resulting in data sets acquired with stable conditions, unstable conditions, and a mix of the two. The error in aggregation of H was small ($>10 \text{ W m}^{-2}$) for DOY 209 when site variability was low during the early-monsoon season (Figure 7). Error was larger (nearly 20 W m^{-2}) for DOY 221 during the early morning, when the site was characterized by a mix of stable and unstable conditions.

For moderate resolution (6.0 m) NS001 data, error in aggregation associated with the stability correction was small ($< 6 \text{ W m}^{-2}$) for all METFLUX sites. As with the yoke-based data, the data were acquired during mid-day when unstable conditions dominated the site ($T_k - T_a > 0$) for most pixels. Results were similar for the coarse resolution (120 m) TM data. The error associated with aggregation of data acquired during the dry season was negligible for both the WGEW and RIP sites. For data acquired during the wet season (DOY 252) at WGEW and RIP, there was greater error though it was still small ($< 5 \text{ W m}^{-2}$).

In conclusion, error associated with aggregation of H when z_o was held constant during pre-aggregation and post-aggregation computations was larger for sites with a mixture of stable and unstable conditions (e.g., morning measurements with thermal scanner and yoke-based measurements at Lucky Hills, DOY 252) than for sites dominated by unstable conditions (mid-day measurements in most cases). The error appeared to decrease with increasingly coarser resolution. The differences in r_a' associated with differences in stability

were especially large when the site was composed of an even mixture of stable and unstable conditions at the pixel scale, due to the high non-linearity of the relation between H and $T_k - T_a$ (e.g., Figure 8). Thus, there were large errors observed for morning measurements at WGEW, when this condition was most common.

Variability in Aggregated Values of H due to Variability in z_o

To test the sensitivity of aggregation of H values to the non-linearity in the relation between $T_k - T_a$ and H due to varying surface roughness conditions, we assumed that z_o was one value for shrub-dominated sites (0.04 m) and another for grass-dominated sites (0.01 m) at the pre-aggregation scale (z_o values computed by Kustas et al., 1994a), and z_o was either 0.04 m or 0.01 m for the post-aggregation scale. Thus, there were three outputs from the aggregation : 1) H computed with a variable z_o value related to shrub or grass cover; 2) H computed with a constant $z_o=0.04\text{m}$; and 3) H computed with a constant $z_o=0.01\text{m}$. This test was conducted by combining the data extracted for each METFLUX site from the thermal scanner and NS001 images. In this way, we could investigate the variability in H associated with all the above-mentioned surface variability and variability in z_o .

For thermal scanner data with fine resolution (0.3 m), there were large differences (nearly 50 W m^{-2}) between values of H aggregated with differing z_o values, especially in mid-afternoon (Figure 9). There was negligible error ($< 5 \text{ W m}^{-2}$) associated with measurements in the early morning when values of H were small ($\approx 35 \text{ W m}^{-2}$). For NS001

data with moderate resolution (6.0 m), the error in aggregation of H was smaller ($< 25 \text{ W m}^{-2}$) (Figure 10) than for the higher-resolution thermal scanner data.

We concluded that substantial errors in H could be expected (at all spatial resolutions) in aggregation of sites with significantly different values of z_o . This was especially true for measurements at midday when the range and magnitude of $T_k - T_a$ values were large. The differences in H associated with differences in z_o could be especially large when values of $T_k - T_a$ were large (e.g., Figure 8). Thus, errors were smaller for morning measurements than for measurements in the afternoon.

It is apparent from the simulated data in Figure 8 that the inclusion of a variable kB^{-1} term in Eq. [10] resulted in a non-linear relation between $T_k - T_a$ and H in Eq. [11]. Though the use of a constant kB^{-1} term would minimize the errors in aggregation of H for heterogeneous sites, it would substantially increase the absolute error in estimation of H (Moran et al., 1994c).

Since λE is computed as a residual in most remote sensing models utilizing Eq. [1] and since we assumed here that there would be little error associated with aggregation of R_n and G , the error associated with aggregation of H is directly related to the error in λE . In some cases (e.g., Figure 9), the error in H increased with its magnitude. Since λE decreases in magnitude as H increases, the resultant error in λE relative to its magnitude would be quite large.

CONCLUSIONS

Previous studies have shown that simple linear aggregation of surface reflectance, albedo and spectral vegetation indices results in near-negligible error for most conditions (Humes and Sorooshian, 1994; Hall et al., 1992). In this study, we confirmed that non-linearities in the TM thermal sensor calibration and conversion from T_r to T_k (accounting for emissivity) were also negligible over most ranges of T_r . However, we found that aggregation of surface energy balance components was more complicated and, under many conditions, the error was substantial.

This study emphasized the need to account for site heterogeneity in selection of a scheme for aggregation of surface energy balance components. This is particularly true for sparsely-vegetated sites, such as the Walnut Gulch semi-arid rangeland. However, the quantification of site heterogeneity is not an easy task since it can be related not only to vegetation cover but also to time of day, time of year and sensor spatial resolution. Thus, it is important to understand the sources of non-linearity in the computation of H and λE in order to understand the errors associated with aggregation.

Caution should be used in aggregation of energy balance components over heterogeneous landscapes with sparse or mixed vegetation types. Substantial error can be obtained with the following conditions:

- Sites which are composed of a mix of stable and unstable conditions;
- Sites which have considerable variations in aerodynamic roughness, especially for highly unstable conditions where $T_k - T_a$ is large; and

- Sites which are characterized by patch vegetation, where the pixel resolution is less than or nearly-equal to the diameter of the vegetation "element" (in most cases, the diameter of the dominant vegetation type or vegetation patch).

Future work should address two issues. First, this analysis addressed only scale issues related to surface heterogeneity. It did not address feedbacks and integrating effects of the atmosphere (Brutsaert, 1986; Jacobs and de Bruin, 1992) that could decrease the potential for error in aggregation that was indirectly illustrated in Figure 8. Second, results from this analysis showed that there was substantial error in aggregation of H for sites with differences in surface roughnesses. These results were obtained with the aggregation of sites where roughness varied by only 0.03 m. The aggregation error for sites with greater variation in roughness lengths, such as mixtures of grassland and forest, could be many times larger than that presented here.

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FIGURES AND TABLES

Figure 1. An example of a scattergram of values of $(T_k - T_a)$ and SAVI measured in well-watered and water-deficit alfalfa treatment plots with differing vegetation cover ranging from near-zero to 100% cover (From Moran et al., 1994a). The solid lines represent the trapezoidal shape that would result from the relation between the SAVI and $(T_k - T_a)$ computed with the Penman-Monteith equation for complete vegetation cover (SAVI=0.8) and bare soil (SAVI=0.1).

Figure 2. Landsat TM image covering WGEW and the area immediately surrounding. The solid lines delimit the boundaries of the WGEW and RIP targets. The flight paths of the Cessna and C130 aircraft intercepted the locations of the 8 METFLUX sites (X).

Figure 3. An illustration of the variability of the yoke-based (0.3 m resolution) and TM spectral data (120 m resolution) for WGEW and RIP on DOYs 156 and 252.

Figure 4. Comparisons of histograms of thermal scanner data to illustrate the variability of $T_k - T_a$ values due to influences of a) time of day, b) time of year, c) vegetation type/cover, and d) sensor spatial resolution.

Figure 5. An investigation of the effect of emissivity and sensor calibration on aggregation of surface temperature estimates from Landsat TM data. Kinetic temperatures (T_k) were calculated using Eqs. [7], [8] and [9] with TM6 radiance and TM estimates of NDVI for WGEW and RIP on DOYs 156 and 252.

Figure 6. An investigation of the effect of the r_s' stability correction on aggregation of sensible heat flux (H) using yoke-based data. Values of H were calculated using Eqs. [7] and [9] - [12], and yoke-based measurements of T_r and NDVI for Kendall (K) and Lucky Hills (LH) on DOYs 156 and 252 at 10:30. z_o was assumed to be constant at both scales, though different for each site.

Figure 7. An investigation of the effect of the r_s' stability correction on aggregation of sensible heat flux (H) using thermal scanner data. Values of H were calculated using Eqs. [10] - [12] and thermal scanner measurements of T_k for Kendall (K) and Lucky Hills (LH) on DOY 209 and DOY 221. z_o was assumed to be constant at both scales, though different for each site.

Figure 8. Comparison of H values computed for a continuum of ($T_k - T_a$) values DOY 252 at 10:30 at Kendall ($z_o = 0.01$ m) and Lucky Hills ($z_o = 0.04$ m) for stable and unstable conditions. These data illustrate the non-linearity of the relation between ($T_k - T_a$) and H, and the differences in the relation due to variations in roughness.

Figure 9. An investigation of the effect of stability corrections of r_a' and site-specific differences in z_o on estimates of sensible heat flux (H) using thermal scanner data. Values of H were calculated using Eqs. [10] - [12] and thermal scanner measurements of T_k for a combined data set of Lucky Hills and Kendall on DOY 209 at 10:12, DOY 209 at 14:40, and DOY 221 at 7:30. z_o was assumed to 0.01 m at Kendall and 0.04 m at Lucky Hills for the bars labeled " z_o measured".

Figure 10. An investigation of the effect of stability corrections of r_a' and site-specific differences in z_o on estimates of sensible heat flux (H) using NS001 data. Values of H were calculated using Eqs. [10] - [12] and NS001 measurements of T_k for a combined set of data from all MF sites on DOY 221 at 10:30. z_o was assumed to be 0.01 m at MF3-MF7 and 0.04 m at MF1, MF2 and MF8 for the bar labeled " z_o measured".

Table 1. General description of instruments deployed during the Monsoon'90 Experiment.

Table 2. Summary of statistics associated with histograms of thermal scanner data (acquired with the Cessna aircraft flying at 0.09 km to 3.35 km above ground level) presented in Figure 4. "Skew" is the 3rd moment about the mean, where a larger absolute value represents a more pronounced skew and a negative number indicates a skew to the

right. "Std. Dev." is one standard deviation from the mean, where $n=143,112$ in all cases.

Table 1. General description of instruments deployed during the Monsoon'90 Experiment.

PLAT-FORM	INSTRUMENT	NO. OF BANDS	WAVE-LENGTH RANGE	TARGETS	FLIGHT DATES (DAY OF YEAR 1990)
Yoke	Exotech Radiometer, IFOV: 15°	4	0.50-0.89 μm	Kendall: 0.48x0.12 km	156, 252
	Everest IRT, IFOV: 15° Footprint: 0.3 m	1	8-13 μm	Lucky Hills: 0.12 x 0.12 km	
Cessna	Thermal Infrared Scanner: IFOV 2.4 mrad Footprint: 0.2 m at 0.09 km AGL 1.7 m at 0.92 km AGL 6.0 m at 3.35 km AGL	1	8-12 μm	Kendall and Lucky Hills: 380 pixels by 480 lines	209, 221 @ 0.09 km AGL 222 @ 0.92 km 209 @ 3.35 km
C-130	NS001 Thematic Mapper Simulator, IFOV: 2.5 mrad Footprint: 6.0 m	8	0.46-12.3 μm	All 8 METFLUX Sites: approximate size 380 pixels by 480 lines	221
Landsat-5	Thematic Mapper (TM) Footprint: 30 m Reflected and 120 m Emitted	7	0.45-12.5 μm	Two targets: WGEW and RIP (Riparian site) 10 km by 3 km each	156, 252

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Figure	Time of Day	Day of Year	Location	Spatial Resolu- tion	Range of $T_r - T_a$ values °C	Skew	Std. Dev.
4a	7:14	221	L. Hills	0.2 m	10.5	0.06	1.44
4a	9:02	221	L. Hills	0.2 m	13.9	-0.73	1.48
4a	7:07	221	Kendall	0.2 m	7.6	-0.45	0.72
4a	9:24	221	Kendall	0.2 m	16.4	-0.57	2.00
4b	9:59	209	Kendall	0.2 m	21.3	-0.75	2.07
4b	9:24	221	Kendall	0.2 m	16.4	-0.57	2.00
4c	14:40	209	Kendall	0.2 m	27.0	-0.85	3.44
4c	14:34	209	L. Hills	0.2 m	25.7	-0.72	4.73
4d	14:40	209	Kendall	0.2 m	27.0	-0.85	3.44
4d	14:51	209	Kendall	6.0 m	13.0	-0.23	1.58

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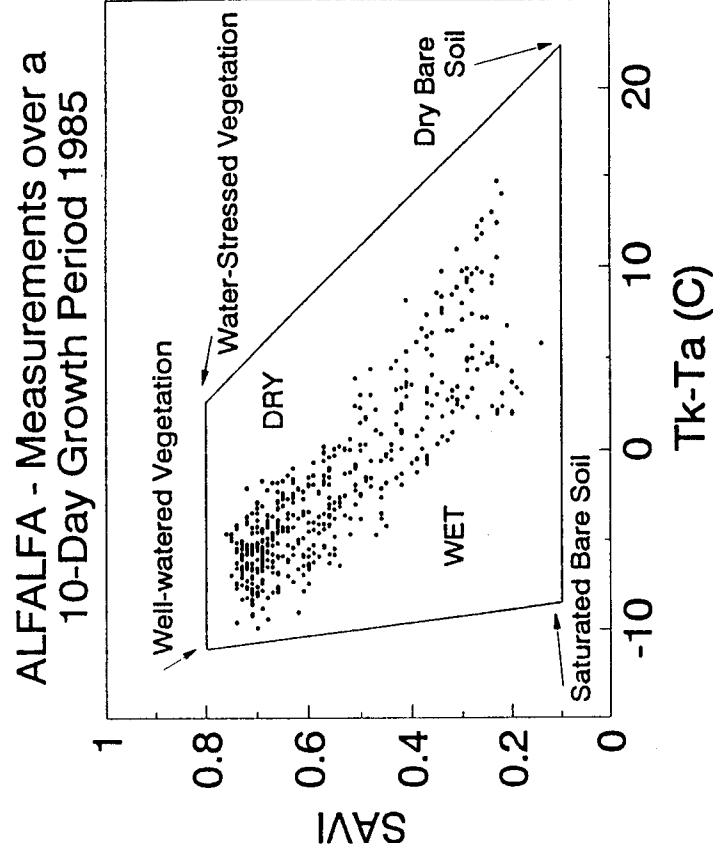


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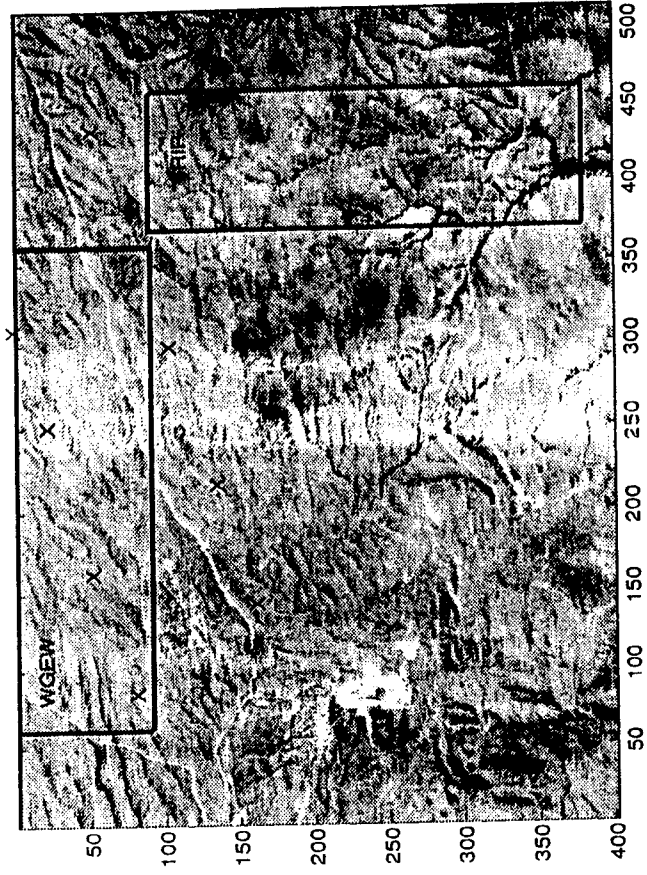


Figure 3. An illustration of the variability of the yoke-based (0.3 m resolution) and TM spectral data (120 m resolution) for WGEW and RIP on DOYs 156 and 252.

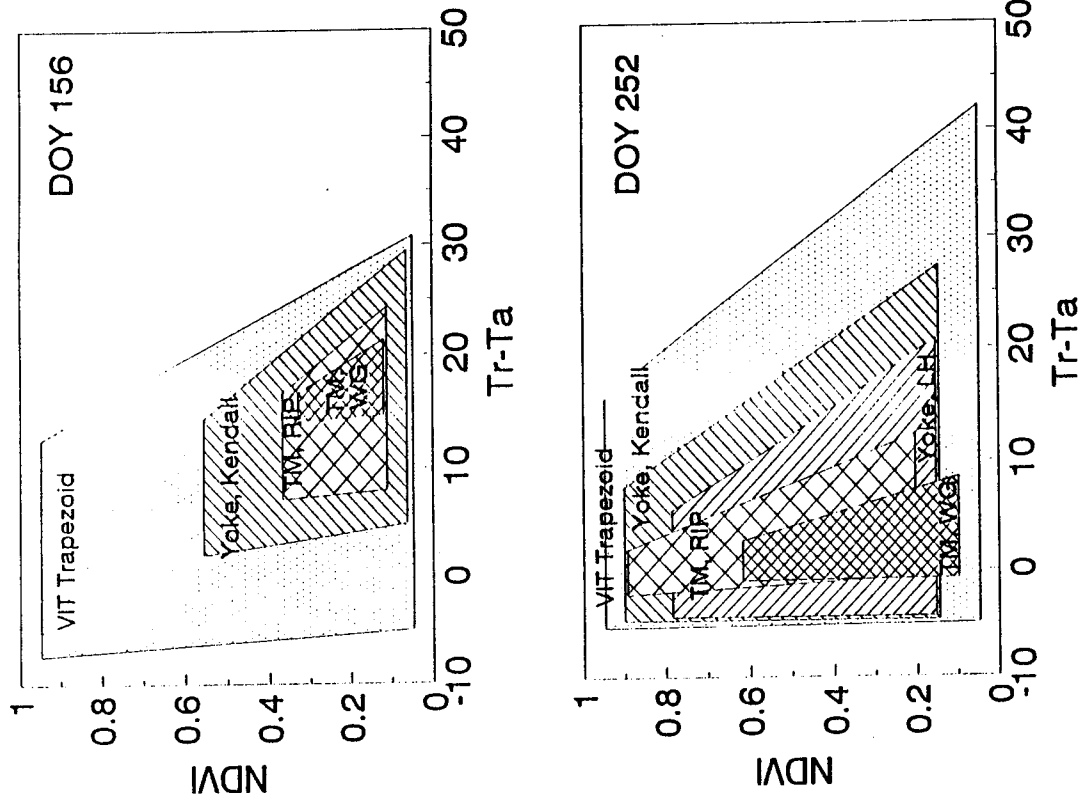


Figure 4. Comparisons of histograms of thermal scanner data to illustrate the variability of T_r - T_a values due to influences of a) time of day, b) time of year, c) vegetation type/cover, and d) sensor spatial resolution.

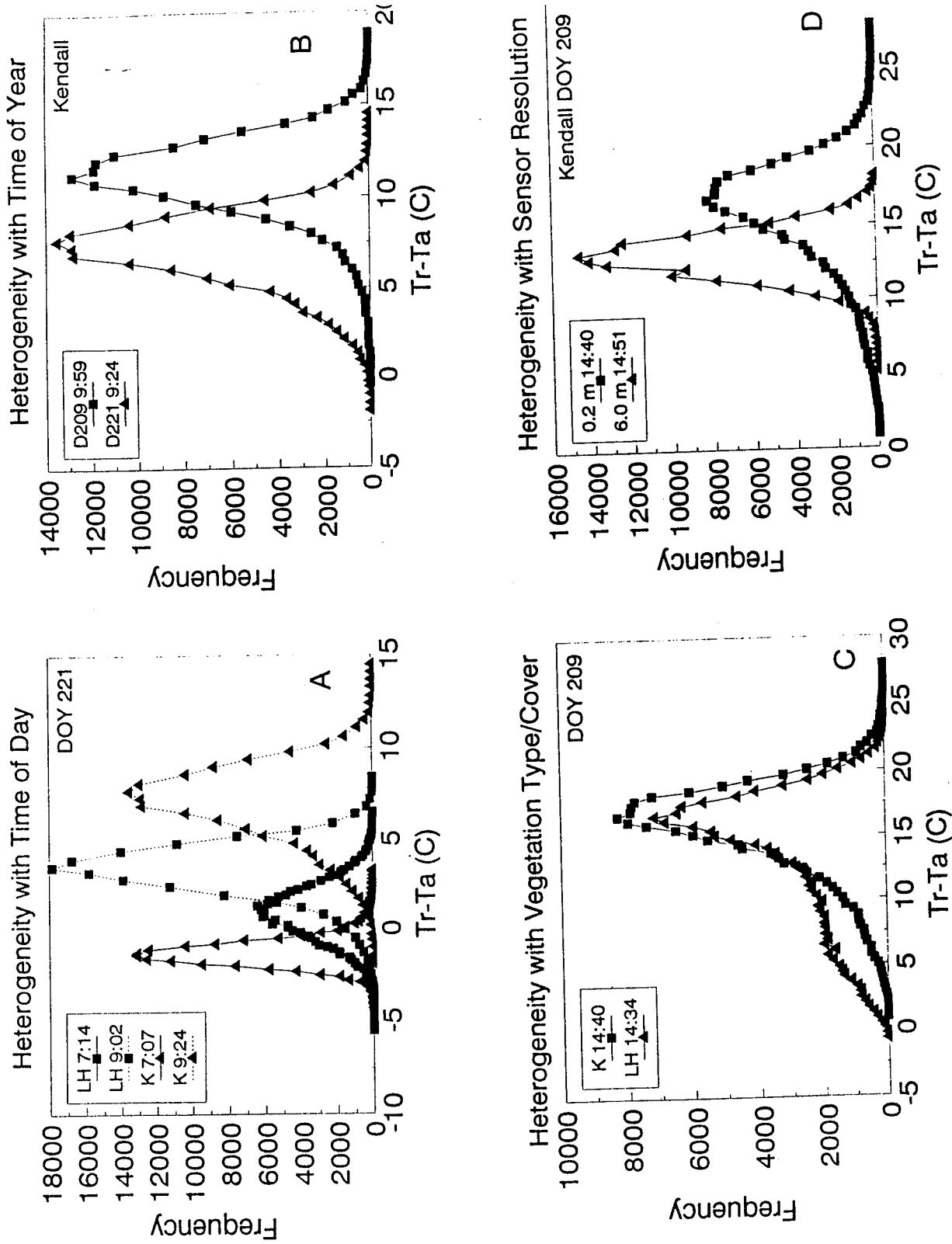


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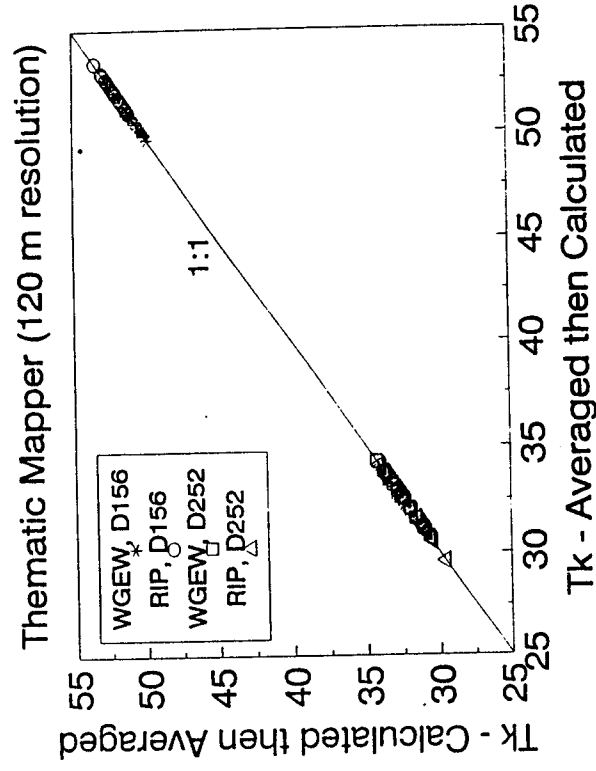


Figure 6. An investigation of the effect of the r_s' stability correction on aggregation of sensible heat flux (H) using yoke-based data. Values of H were calculated using Eqs. [7] and [9] - [12], and yoke-based measurements of T_s and NDVI for Kendall (K) and Lucky Hills (LH) on DOYs 156 and 252 at 10:30. z_o was assumed to be constant at both scales, though different for each site.

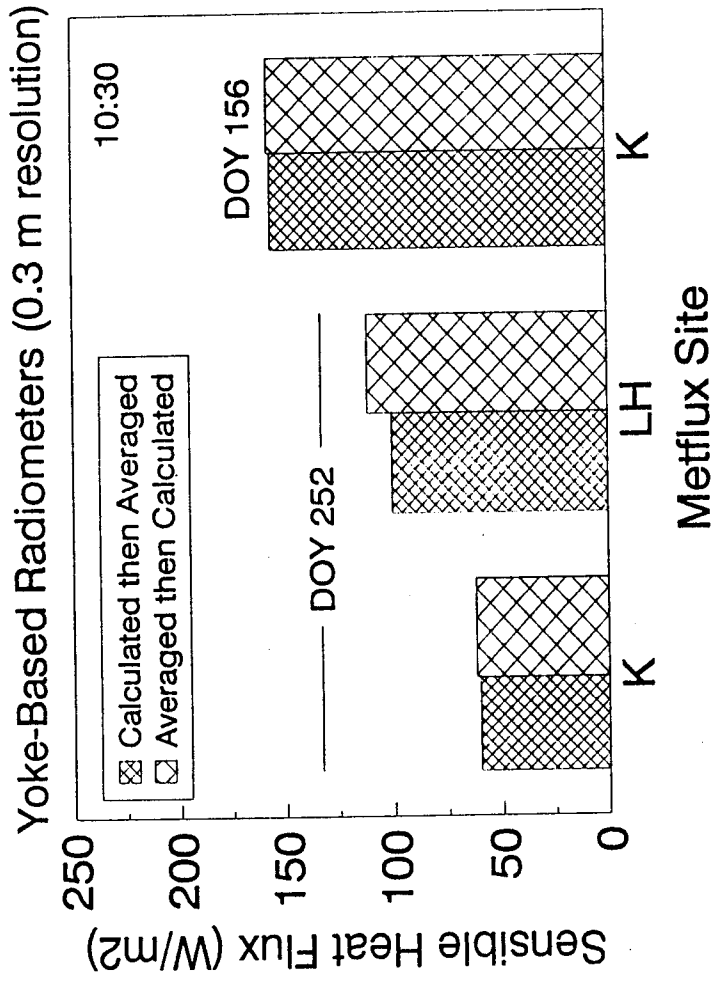


Figure 7. An investigation of the effect of the r_s' stability correction on aggregation of sensible heat flux (H) using thermal scanner data. Values of H were calculated using Eqs. [10] - [12] and thermal scanner measurements of T_k for Kendall (K) and Lucky Hills (LH) on DOY 209 and DOY 221. z was assumed to be constant at both scales, though different for each site.

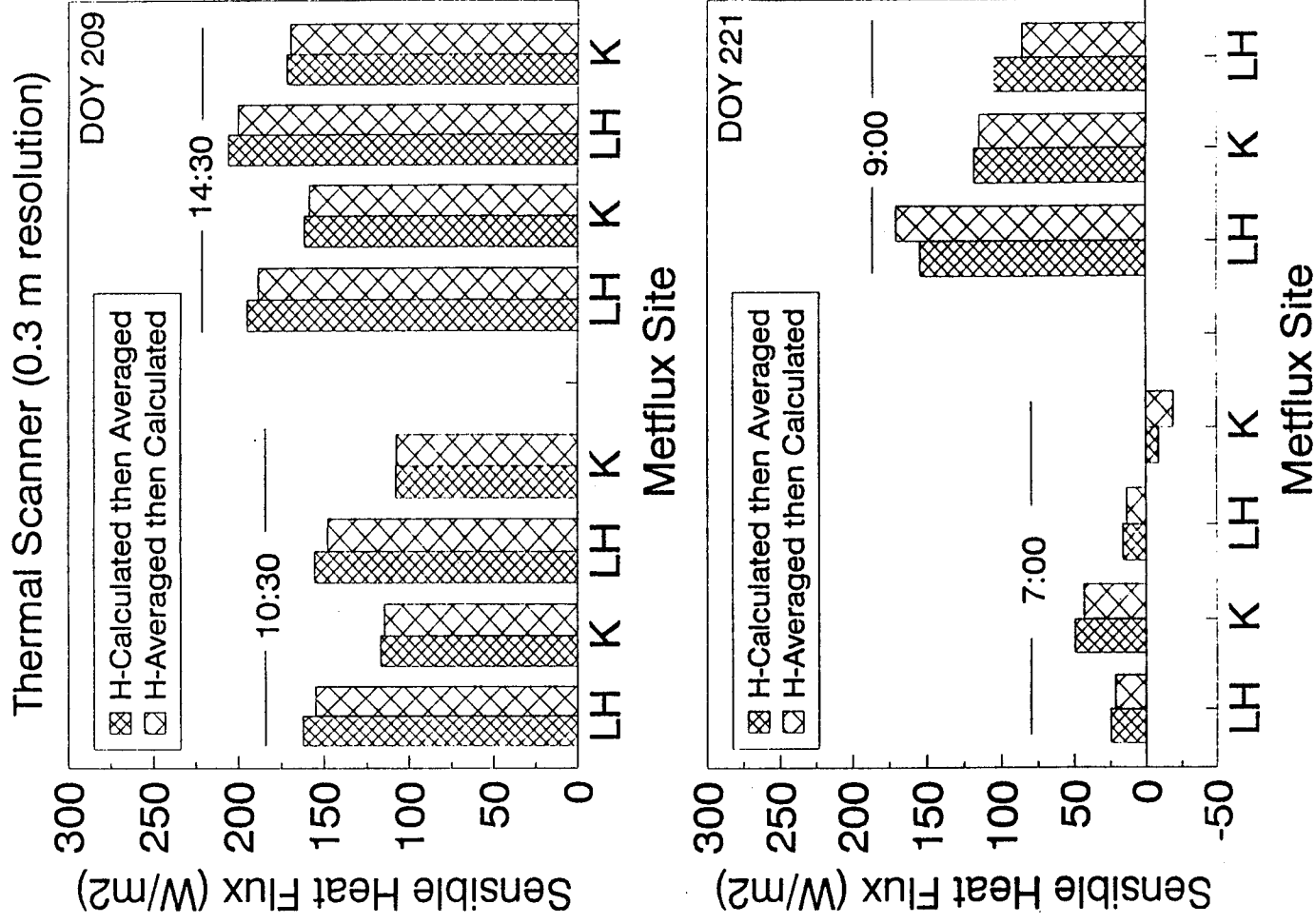


Figure 8. Comparison of H values computed for a continuum of $(T_k - T_a)$ values DOY 252 at 10:30 at Kendall ($z_0=0.01$ m) and Lucky Hills ($z_0=0.04$ m) for stable and unstable conditions. These data illustrate the non-linearity of the relation between $(T_k - T_a)$ and H, and the differences in the relation due to variations in roughness.

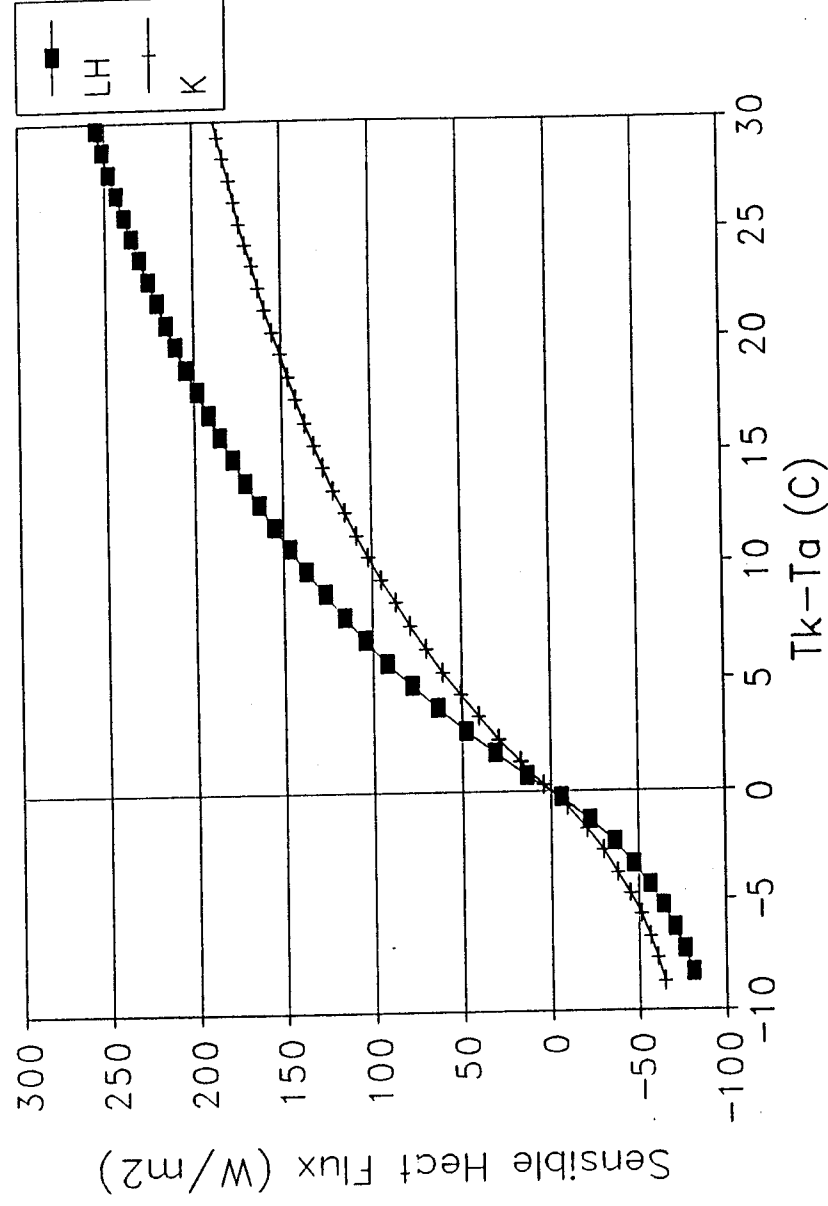


Figure 9. An investigation of the effect of stability corrections of r_s' and site-specific differences in z_0 on estimates of sensible heat flux (H) using thermal scanner data. Values of H were calculated using Eqs. [10] - [12] and thermal scanner measurements of T_s for a combined data set of Lucky Hills and Kendall on DOY 209 at 10:12, DOY 209 at 14:40, and DOY 221 at 7:30. z_0 was assumed to 0.01 m at Kendall and 0.04 m at Lucky Hills for the bars labeled "z₀ measured".

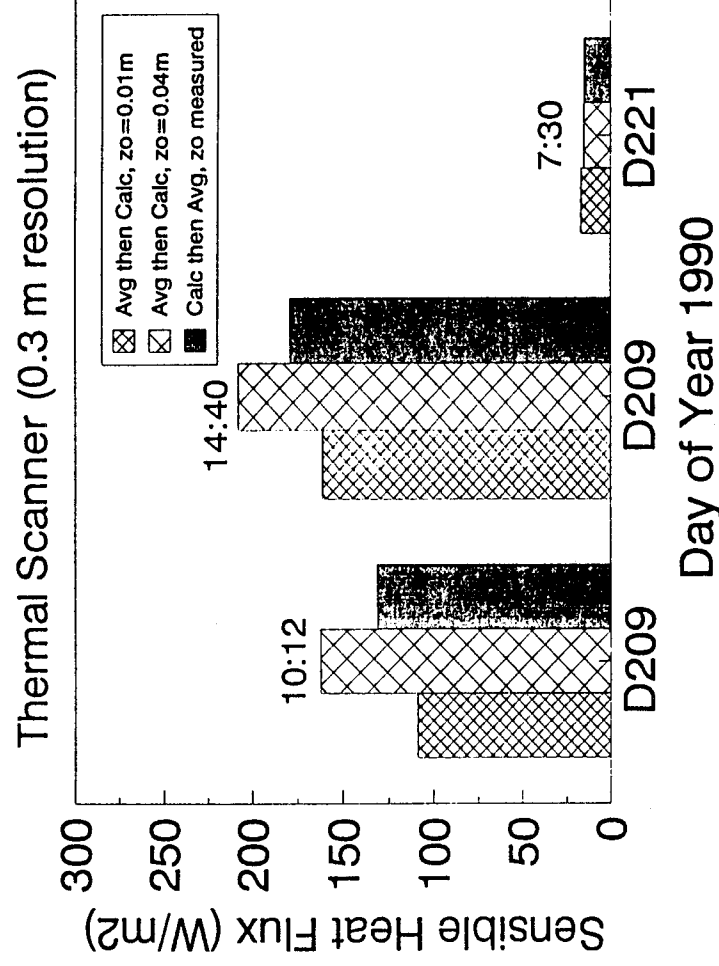
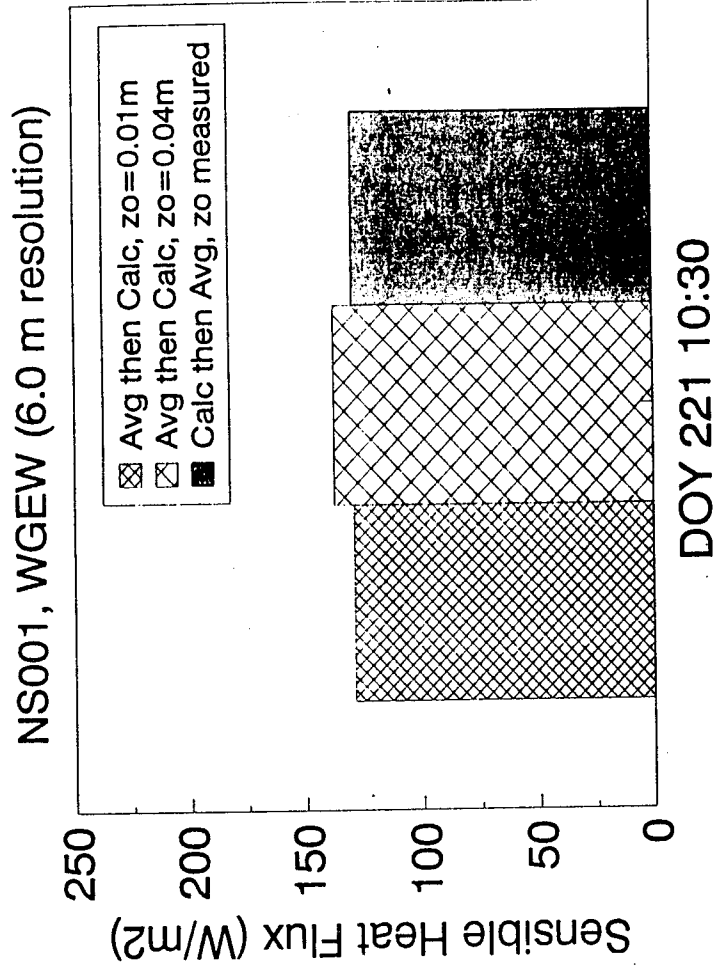


Figure 10. An investigation of the effect of stability corrections of r_s' and site-specific differences in z_o on estimates of sensible heat flux (H) using NS001 data. Values of H were calculated using Eqs. [10] - [12] and NS001 measurements of T_k for a combined set of data from all MF sites on DOY 221 at 10:30. z_o was assumed to be 0.01 m at MF3-MF7 and 0.04 m at MF1, MF2 and MF8 for the bar labeled " z_o measured".



Defining area-average parameters in meteorological models for landsurfaces with mesoscale heterogeneity

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Abstract

After a brief review of existing methods developed to describe the sub-grid variability of soil and vegetation at large scale, the paper summarizes aggregation results obtained in the framework of the Hapex-Mobilhy and EFEDA field experiments. The parameter aggregation method is based on the estimation of effective parameters for the soil and for the vegetation, representative of the large area. In conditions of moderate land-surface variability the effective surface fluxes computed using a single column model forced by effective surface parameters are close to the areally-averaged turbulent fluxes estimated by a mesoscale model. A simple method for accounting the sub-grid variability of intercepted rain is tested.

1 Introduction: a brief review on sub-grid heterogeneity in atmospheric models

The aim of this paper is to address the issue of the description of land-surface heterogeneity within a large area comparable in size with a grid box of a General Circulation Model (GCM). In the ten last years, substantial efforts have been paid to solve this problem through the organization of large field experiments and also with the development of a wide range of surface schemes to be applied in GCMs. Two types of methods have been proposed to describe the subgrid variability of soil and of vegetation in a large area:

- (i) in the flux aggregation method, the surface flux from each land-use category in the gridbox is estimated and then a spatial average is computed to obtain an areally-averaged surface flux.
- (ii) in the parameter aggregation method, it is proposed to describe the subgrid variability of soil and vegetation types through 'effective' parameters which can be used in a single column atmospheric model to compute an effective surface flux. The effective flux is expected to be comparable with the areally-averaged flux, despite the non-linear relationships between surface parameters and surface fluxes.

The paper focusses on the representation of surface fluxes and does not address the parameterization of the mesoscale fluxes which could be generated by the subgrid-scale landscape discontinuities in GCMs. The paper of Pielke (this issue) proposes some investigations on the topic.

After a brief review of existing schemes accounting for land-surface heterogeneity, we concentrated on results obtained with the parameter aggregation method. The upscaling method is based on mesoscale modelling performed in the framework of the Hapex-Mobilhy (1986) and EFEDA (1991) field experiments.

Avisar and Pielke (1989) and Koster and Suarez (1992) proposed a surface parameterization with an explicit description of the subgrid heterogeneity. In these so called 'tile' approaches a surface transfer scheme is applied to each land use and is coupled independently with the GCM first level. Each tile has its own prognostic variables. The authors argued that the tile approach (the LEAF and MOSAIC surface schemes) are to some extent simpler than the classical 'mixture' big leaf model. For instance the resistance network in Leaf is simpler than in SiB. An other advantage is that the tile approach can provide additional information on

the subgrid distribution of state variables (soil water, near-surface atmospheric quantities,...). However, there is no evident superiority of the tile approach on the big leaf concept. Besides the evident difficulty for the validation with observational data, the implementation of the tile approach rises a lot of problems which have not to be underestimated. A first problem concerns the mapping of the vegetation and soil properties with a higher resolution. In fact the required information is never available and one is obliged to prescribe most of the surface parameters needed and to simplify as much as possible the level of subgrid variability described by the scheme. Another problem concerns the description of the subgrid variability of the soil water content. In principle, the subgrid water budget should be calculated for each type of land surface area (mosaic of agricultural fields, forests, built-up areas,...). If the problem of the subgrid variability of rainfall has also to be considered, one measure immediately the complexity of the task. Finally, the simulation of the subgrid variability of soil moisture can also complicate the issue of soil moisture assimilation in numerical weather prediction models. Avisar (1991) introduced the concept of intra-patch heterogeneity referring to small-scale variability of soil and vegetation. He suggested to describe this variability from a statistical-dynamical approach which consists of using probability density functions (pdf) for the various parameters of the soil-plant atmosphere system. In vegetated patches, energy balance equations are solved for an array of leaves that are assumed to have a distribution of positions and of stomatal conductances. But again, the superiority of the statistical-dynamical approach is not clearly demonstrated.

In the parameter aggregation method, it is proposed to overcome the problems of nonlinearity between the surface parameters and the surface fluxes, by the choice of appropriate operator. Several studies have concentrated on upscaling from 100 m to 10 km. Mason (1988) proposed a method for estimating an effective roughness length using the concept of blending height. Dolman (1992) examined the way of combining a distribution of stomatal resistances into an effective resistance. Blyth et al. (1993) studied various methods of averaging aerodynamic and surface resistances and suggested that effective parameters for each type of surface flux should be calculated separately. Concerning the upscaling of soil properties, there are a lot of studies addressing this issue using stochastic modelling (see Kabat's paper in this special issue).

Examples of aggregation for spatial scales ranging from 10 km to 100 km can be found in many GCMs where averaging the input parameters to the resolution of the GCM is done prior to the simulation (Dickinson et al. 1986)

2 Assessing simple aggregation methods based on mesoscale modelling

This section summarizes studies on parameter aggregation based on mesoscale modelling applied to selected days of the field experiments Hapex-Mobilhy86 (Andre et al., 1986) and EFEDA (Bolle et al. 1993). Roughly speaking, the first step of the method was to calibrate a land surface scheme (the ISBA scheme, Noilhan and Planton 1989) against observations of surface fluxes. The second step concerns the mapping of surface parameters within a mesoscale domain encompassing the experimental area (Figure 1). The third point is the validation of the mesoscale model using detailed comparisons with all the available data taken during selected days. Of course, the mesoscale model should use the surface scheme previously calibrated at the local scale. The validation of the mesoscale model is a very important point since the 3D model results will constitute the reference against which to test aggregation methods. The validation has to be performed with the purpose of spatial extension using spatially integrated observations of fluxes. Two examples of comparison of observed and simulated fluxes representative of the Hapex-Mobilhy and EFEDA large areas are given in Figures 2 and 3. Figure 2 refers to the Hapex-Mobilhy experimental square, half covered by a pine forest, the remaining part corresponding to an agricultural area. This diagram is an attempt to synthesize all the observations of surface fluxes available within the large area in order to estimate aggregated fluxes for comparison with areally-averaged fluxes predicted by the mesoscale model (Bougeault et al. 1991, Noilhan et al. 1991). The observed areally-averaged fluxes (solid line in Figure 2) are computed from the observed fluxes over the forest (one single site of observations) and from the average of the fluxes measured at nine flux stations located in the agricultural area. Furthermore, it was assumed that the single measurement point over the forest was representative of the energy balance of the whole coniferous forest and, naturally, that the others flux observations were representative of the non-forested part of the area. The observed averaged fluxes were estimated as the sum of the fluxes from the two contributive areas, weighed by their respective fractional cover. Figure 2 gives the daily course of the sensible, latent and soil heat fluxes. The sensible heat flux was about two times lower in the crop area than in the forest (Noilhan et al. 1991). On the other hand, the evaporation flux was much less scattered. The decrease of the forest transpiration was interpreted as the result of stomatal control due to increasing atmospheric

specific humidity deficit. The areally-averaged fluxes predicted by the mesoscale model are close to the corresponding observed quantities, except maybe for the soil heat flux where a time lag is noted between the predicted and observed averaged fluxes. The other very important source of information for validating the 3D simulations are the data taken by instrumented aircrafts, flew at low levels. For instance, in the Hapex area it has been possible to interpret spatial variation of fluxes in the atmospheric boundary layer in relation with the horizontal variation of surface fluxes. Thanks to a careful calibration of the mesoscale model, the observed pattern of fluxes in the atmosphere were correctly simulated (Noilhan et al. 1991). Figures 3 and 4 give an additional example of comparison between observed and simulated latent heat flux during EFEDA. This example shows that numerical models can be used not only for upscaling purposes but also to interpret results from field experiments. There are cases where the experimental results are ambiguous and very difficult to fully understand, while the simultaneous consideration of output from the corresponding numerical model does greatly improve the understanding of the phenomena. As an example, if one considers the latent heat flux map derived from aircraft observations (Figure 3a), it is difficult to understand why there are so large differences at moderate height (i.e. 400 m asl) over short distances (i.e., less than 100 km between the Barrax and Tomelloso sites). Such a gradient is surprising, especially if one remembers that the surface latent heat flux over the whole region does in fact vary in the opposite way. Quite large surface evaporation was observed in the eastern part, due to irrigation, and significantly reduced latent heat fluxes in the western part, a semi-arid dry area. The explanation can only be found through numerical modelling and in that case to the explicit description of land-sea breeze effects. The 3D mesoscale model implemented over the EFEDA region (Noilhan 1994) simulates the development of an intense land-sea breeze circulation, leading to moistened air aloft in the eastern part, closer to the Mediterranean. This results in totally different latent heat flux vertical profiles over these two locations, as shown in Figure 4. In the Barrax area, the latent heat flux is decreasing with height because entrainment of dry air at the top of the planetary boundary layer is reduced by the advection of moist air. On the other hand, the latent heat flux over the western Tomelloso area is increasing with height, due to the strong entrainment of dry air at the top of the boundary layer. The strong value of entrainment is associated with the rapidly growing planetary boundary layer and the large discontinuity of specific humidity at the top of the boundary layer. Such quite complicated processes can be

clearly exhibited only through numerical mesoscale modelling since no experimental program will ever be able to gather such a wealth of information.

2.1 Aggregation for dry surfaces

After validation, the mesoscale model can be used for generating the horizontal distribution of surface turbulent fluxes of sensible and latent heat over a large area. By summing the individual values over all the model grid points, one has to access to the very likely value of the area-averaged surface fluxes, used as reference values for testing parameter aggregation methods. The method consists to derive 'effective' parameters for the soil texture and the vegetation in order to compute an 'area-averaged' flux that agrees with the average of the fluxes simulated by the 3D mesoscale model (Noilhan and Lacarrere, 1995). The averaging operators for the vegetation parameters are chosen in order to be consistent with an arithmetic averaging of the fluxes themselves. Thus, surface albedos, leaf-area indices and vegetation cover values are averaged linearly. On the other hand, the roughness lengths are averaged logarithmically, while the stomatal-resistances are inversely-averaged. Concerning the aggregation of the soil properties, the method is based on the fact that the soil hydraulics characteristics are mainly determined by the particle size distribution. For instance, the saturation moisture content used in ISBA (Noilhan and Planton 1989) is linearly dependent on the sand content of the soil. A set of continuous relationships between the soil properties and the sand or clay contents have been established. These predictive relationships for the thermal-hydraulic parameters of ISBA can be used when the fractions of clay and sand of the gridbox are known. Noilhan and Lacarrere (1995) discussed the way of areally-averaging values of sand and clay contents in order to estimate effective soil properties. A simple averaging method was tested as a first guess to investigate the aggregation of soil properties. Besides its simplicity, the idea is to conserve the total water holding capacity of the GCM gridbox.

Thanks to the maps of soil and vegetation types derived for the Hapex-Mobilhy and EFEDA modelling programs, the method has been tested in the two domains (Figure 1). In order to compare the two zones in term of vegetation variability, Figure 5 illustrates the distribution of the surface conductances in the mesoscale model applied to the experimental zones. The parameter $g_s = LAI/R_{smin}$ combines the leaf-area index and the minimum surface resistance prescribed to each grid box of the mesoscale model using the vegetation classifications and 1D

calibration tests with ISBA. The bi-modal shape of the pdf of surface conductance for the Hapex area characterizes the partitioning of the area into the pine forest and the agricultural zone. On the other hand, the pdf is more regular for the EFEDA area, with the exception of high values of surface conductances associated to irrigated crops.

The results of the aggregation are given in Figure 6 for the latent heat flux. The effective flux computed with the 1d model is in fair agreement with the reference 3D flux for the two zones. However, one should note that the scatter is rather small in the mesoscale model. The dotted-line in Figure 6a corresponds to the latent heat flux simulated by the 1D model fed by parameters corresponding to the dominant vegetation type. Values computed this way, clearly underestimate the averaged value of the latent heat flux. This is not a surprise, as subgrid scale variability is not at all taken into account in this oversimplified approach. The parameter aggregation method works satisfactorily also for the others terms of the surface energy balance (not shown), with the exception of the bare soil evaporation component in the case of large variability of soil texture (e.g., juxtaposition of sandy and loamy soils in the Hapex experimental area). Indeed, the effects of non-linearity seem more pronounced in the cases of soil properties than for vegetation.

2.2 Subgrid precipitation and interception loss

Subgrid precipitations and its relation with interception loss is an important issue in GCM modelling. It is well known that the partitioning of evaporation between wet and dry canopy is incorrect if the precipitation is uniformly spread in the grid box. Parameterizations of the convection in GCM do not provide the spatial distribution of rain within the grid. Shuttleworth proposed to describe this variability using a probability distribution of rainfall assuming that a fraction μ of the grid is covered by the convective rain. Inclusion of the pdf in an interception scheme provides a subgrid parameterization for through fall of intercepted rain. This assumption has been examined by Pitman et al. (1990) and by Dolman and Gregory (1992) using a single column model. Pitman et al. noted a large sensitivity of evaporation of intercepted water to the fractional area μ and Dolman and Gregory showed how this factor affects the partitioning between dry and wet evaporation. More recently, the Shuttleworth's proposal was tested within a GCM by Lean and Rowntree (1993) and by Mahfouf et al. (1995), among others. Both studies showed a great improvement of the simulation of interception losses in the Amazon basin. The

method to estimate the parameter μ from various quantities (rainfall intensity, GCM grid mesh and time step) has to be defined. In order to simplify the approach, a constant value of 0.1 is often used.

Mesoscale models also offers some potential for investigating the subgrid representation of interception. Aggregation of surface fluxes from partially wet areas at mesoscale was studied by Blyth et al. (1993). In this study, the Peridot mesoscale model was used to model a frontal intrusion in south west France during Hapex-Mobilhy. The front crossed the area during the day, producing a small amount of rainfall (10 mm). A picture of the spatial distribution of intercepted rainfall simulated by the model is shown in Figure 7. Around midday, the northern part of the domain corresponded to dry areas while significant intercepted rain was simulated in the southern part. Blyth et al. 1993 showed that the model was able to reproduce the salient features of this mesoscale event. The results of the simulation were subsequently used to investigate the aggregation problem of intercepted rain. This was achieved by comparing the results of the three-dimensional model where the heterogeneity is explicitly modelled at the 10 km scale with the results of a simple one-dimensional model, where the heterogeneity is allowed for within the surface scheme. A mesoscale sub-domain was chosen, encompassing the Landes forest which has a rather large interception capacity. From the mesoscale simulation, it was possible to compute the fractional area μ covered by rain within the large domain (Figure 8). In the mesoscale model, it is assumed that uniform rain covers the grid square. This assumption does not hold for large scale models however and it is necessary to account for sub grid variability of rainfall distribution. In the 1D model the calculation of the fraction of vegetation covered by intercepted rainfall was multiplied by the values of μ computed from the 3D model. Figure 9 shows the diurnal cycle of the area-average evaporation of intercepted precipitation from the 3D model and from the 1D model using various methods of aggregation. The large standard deviation of evaporation in the 3D model is due to the fact that a fraction of the domain is dry (evaporation of intercepted rain equals zero) while the remaining wet part evaporates at the potential rate. If the sub-grid distribution of rain is not taken into account in the 1D model, the evaporation flux is clearly overestimated because the interception reservoir is maintained at its saturated value. The reduction of the fractional wetted area has the effect of bringing the rate of evaporation down to levels given by the 3D model. Additionally, Figure 9 shows that the 'GCM' effective evaporation of intercepted rain is significantly improved if the daily cycle of μ

is applied in the aggregation scheme, instead of using a constant value as usually done in many GCMs

3 Conclusions

Two possibilities can be used to describe the subgrid variability of land-surface in GCMs: to describe explicitly the exchanges for an arbitrary number of vegetation and soil types which co-exist in a grid square or to define effective parameters characterizing the subgrid-scale variability. This second approach has been discussed from aggregation studies performed in the framework of the Hapex-Mobilhy and EFEDA modelling programs. The aggregation method are tested by comparing effective surface fluxes computed with a single column model with areally-averaged fluxes estimated from mesoscale modelling. After a careful validation of the mesoscale simulation, the method has been tested for selected clear days and for a rainy day. Despite the non-linear relations between surface fluxes and surface parameters, it is found that the effective surface fluxes match the area-averaged 3D fluxes with a small relative error. The aggregation studies are based on real cases for which moderated land surface variability of soil water was observed. Finally, it was shown that subgrid-scale interception of rainfall by the vegetation can be simply accounted for provided that the fractional area covered by the rainfall is known. In the case of 5 June 1986, the diurnal variation of the fractional area was simulated by a mesoscale model.

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Figure 2: Areally-averaged surface fluxes for the Hapex-Mobilhy area on 16 June 1986: (a) sensible heat flux (b) latent heat flux (c) soil heat flux. Daily variation of observations at the forest site (open circle) and at nine flux stations in the crop area (full circle for average and vertical bars for standard deviation). The observed aggregated fluxes (continuous line) are computed from the observations at the forest site and at the nine stations. The fractional areas used in the averaging are 0.56 for non-forested and 0.44 for the area occupied by the pine forest. The predicted aggregated fluxes (dashed line) are averages of the fluxes computed by the mesoscale model in the Hapex area (e.g. 91 model grid boxes)

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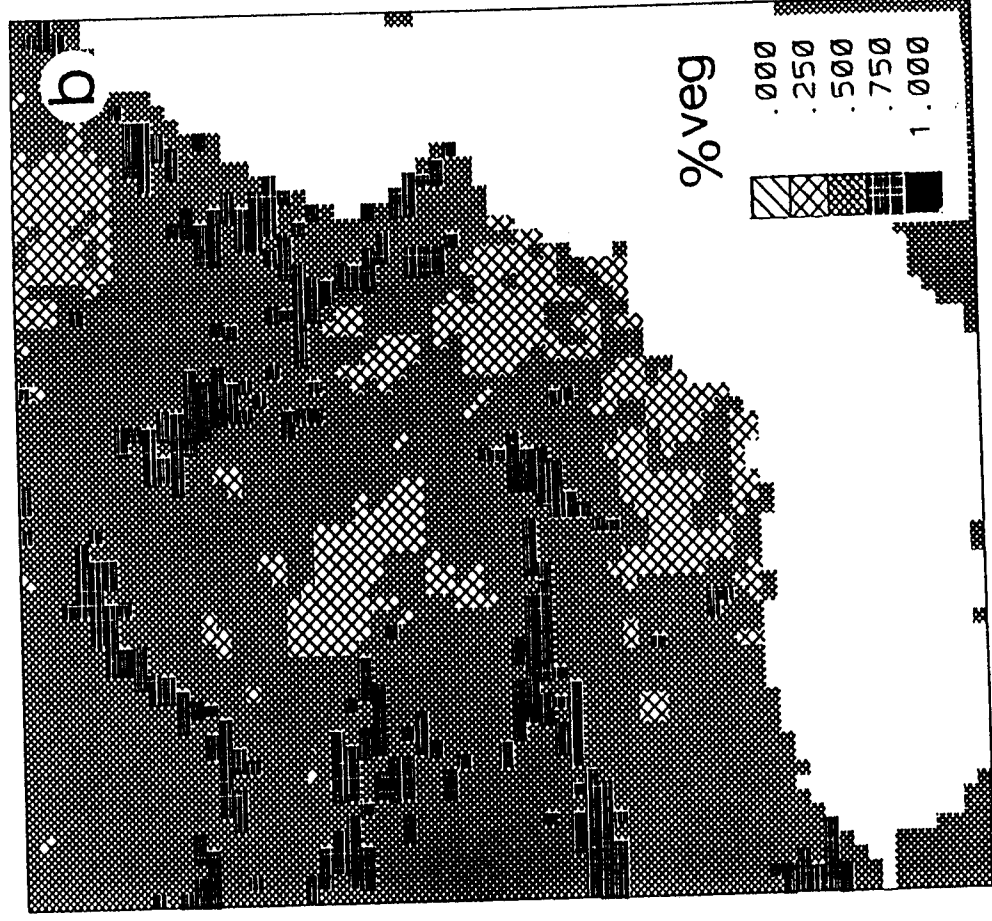
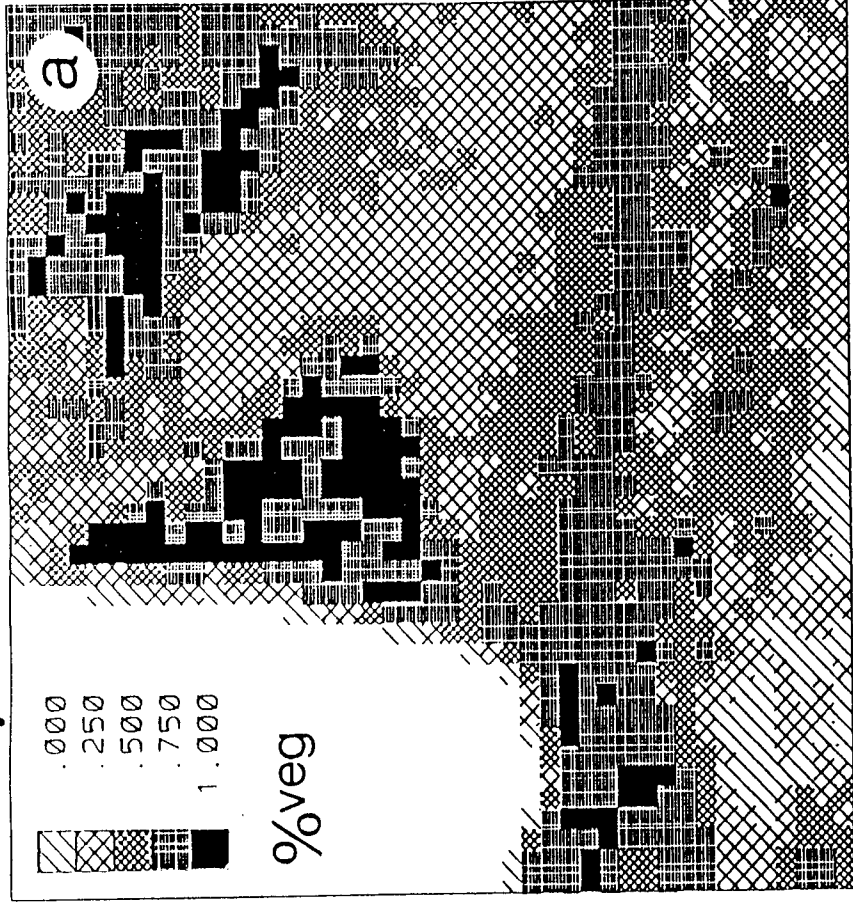
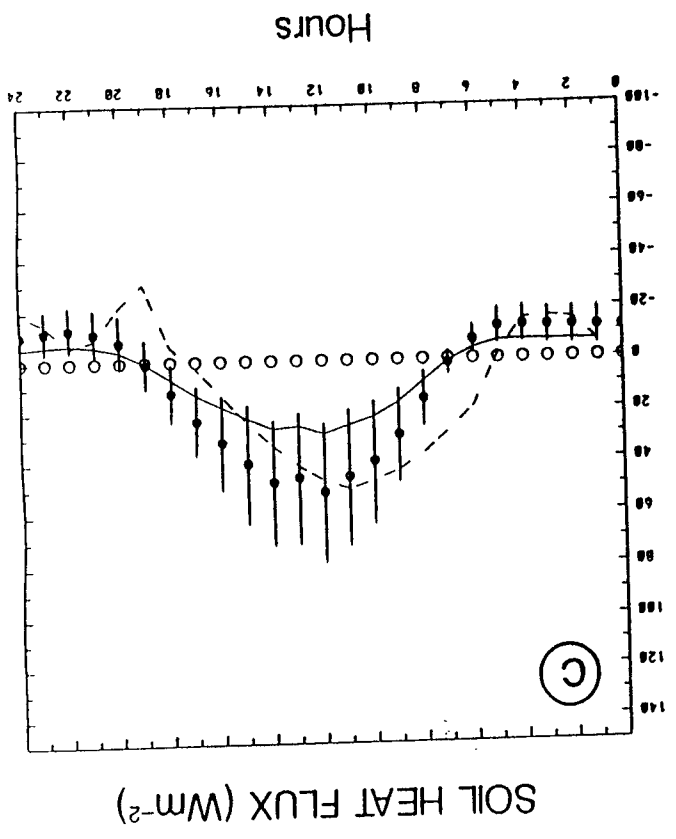
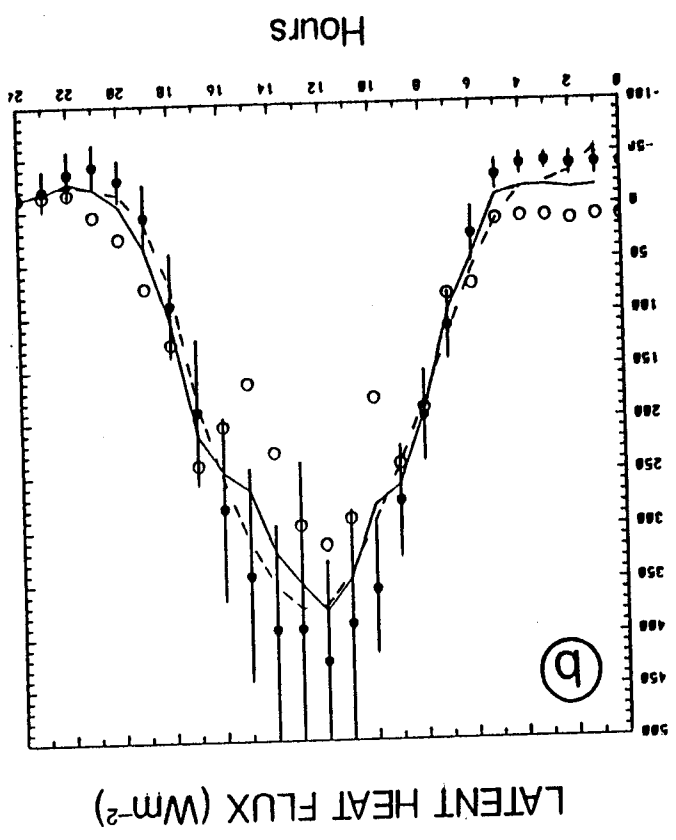
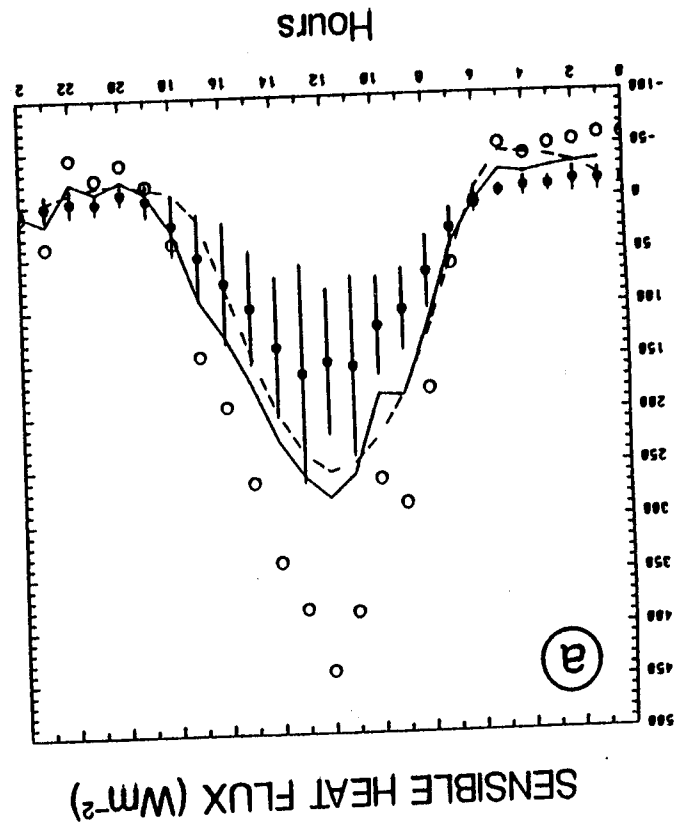
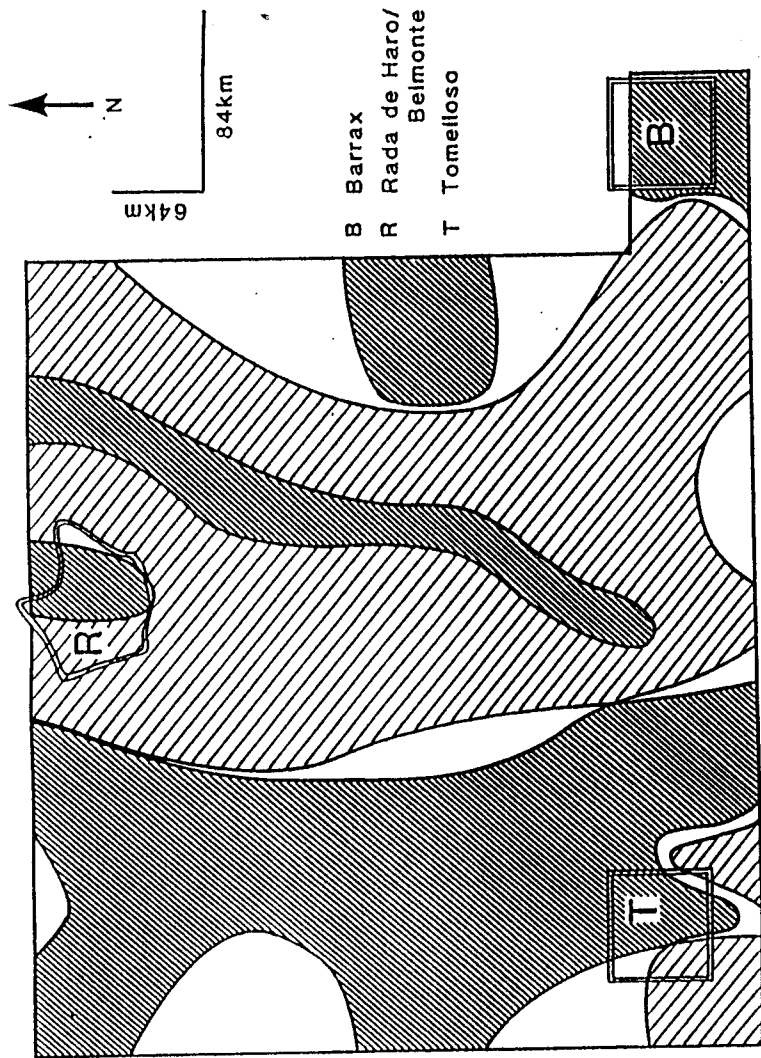


Figure 2



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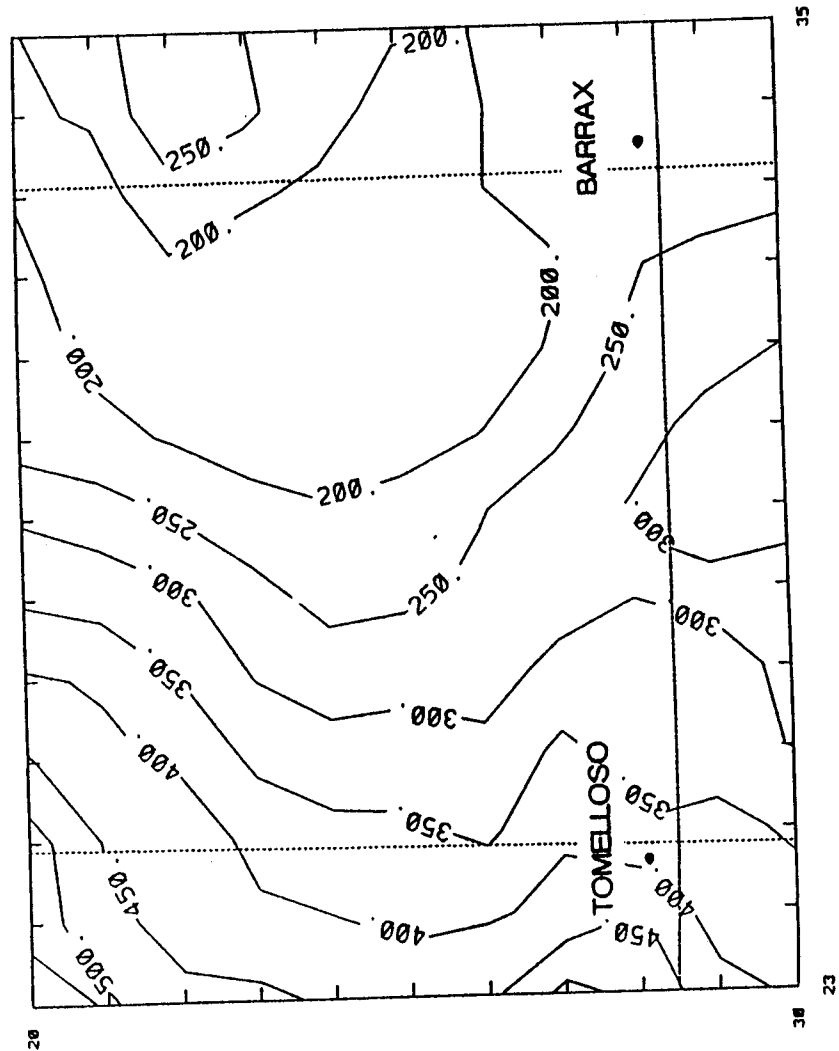


Figure 3

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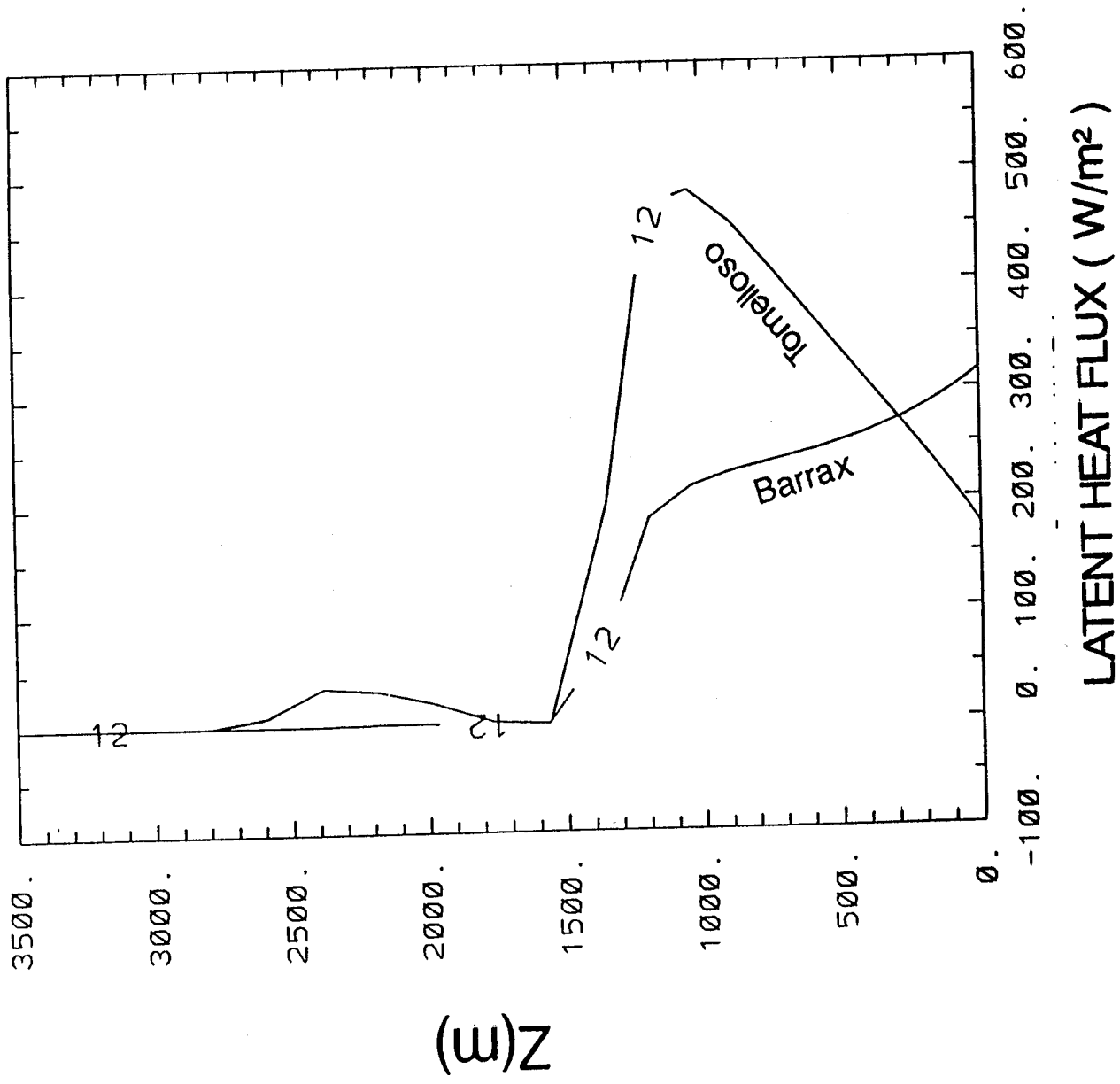
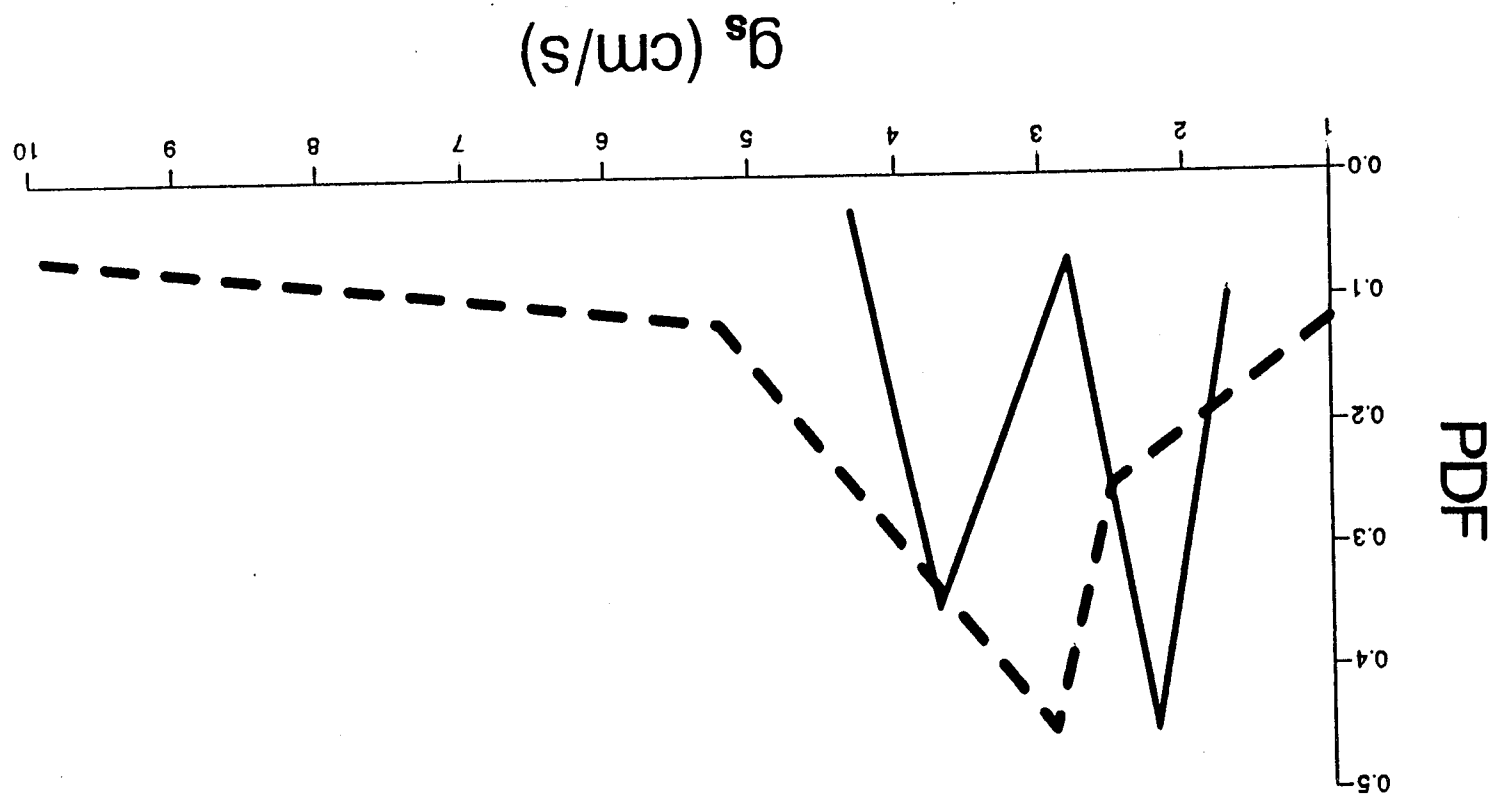
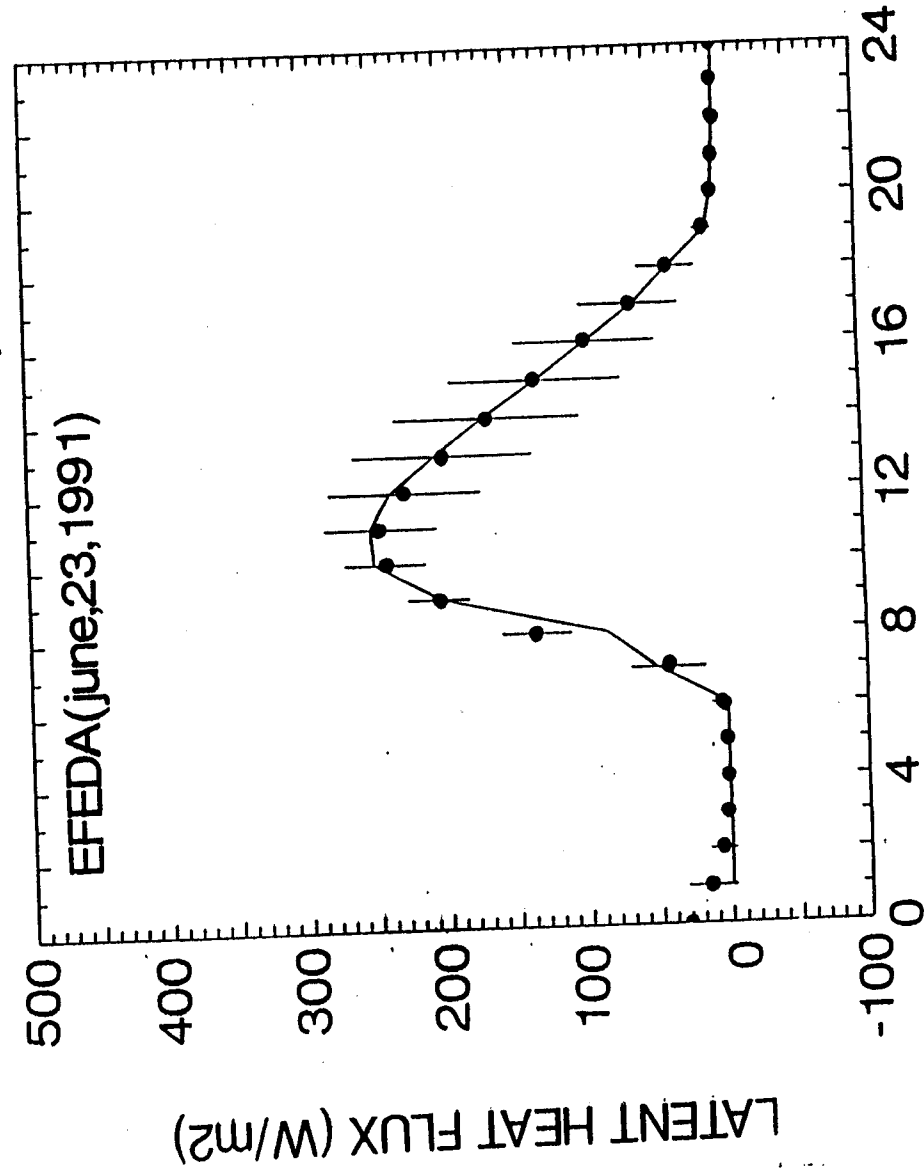
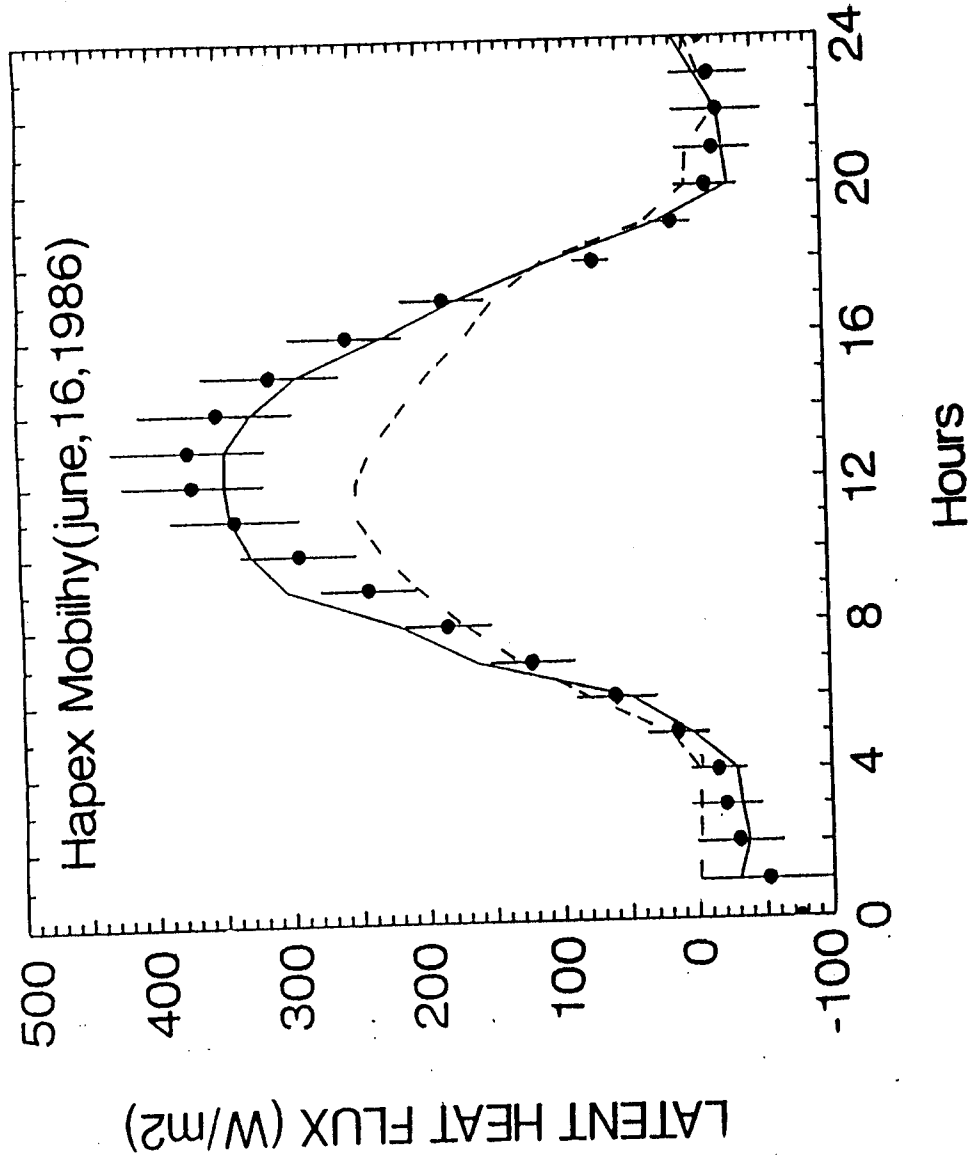


Figure 4





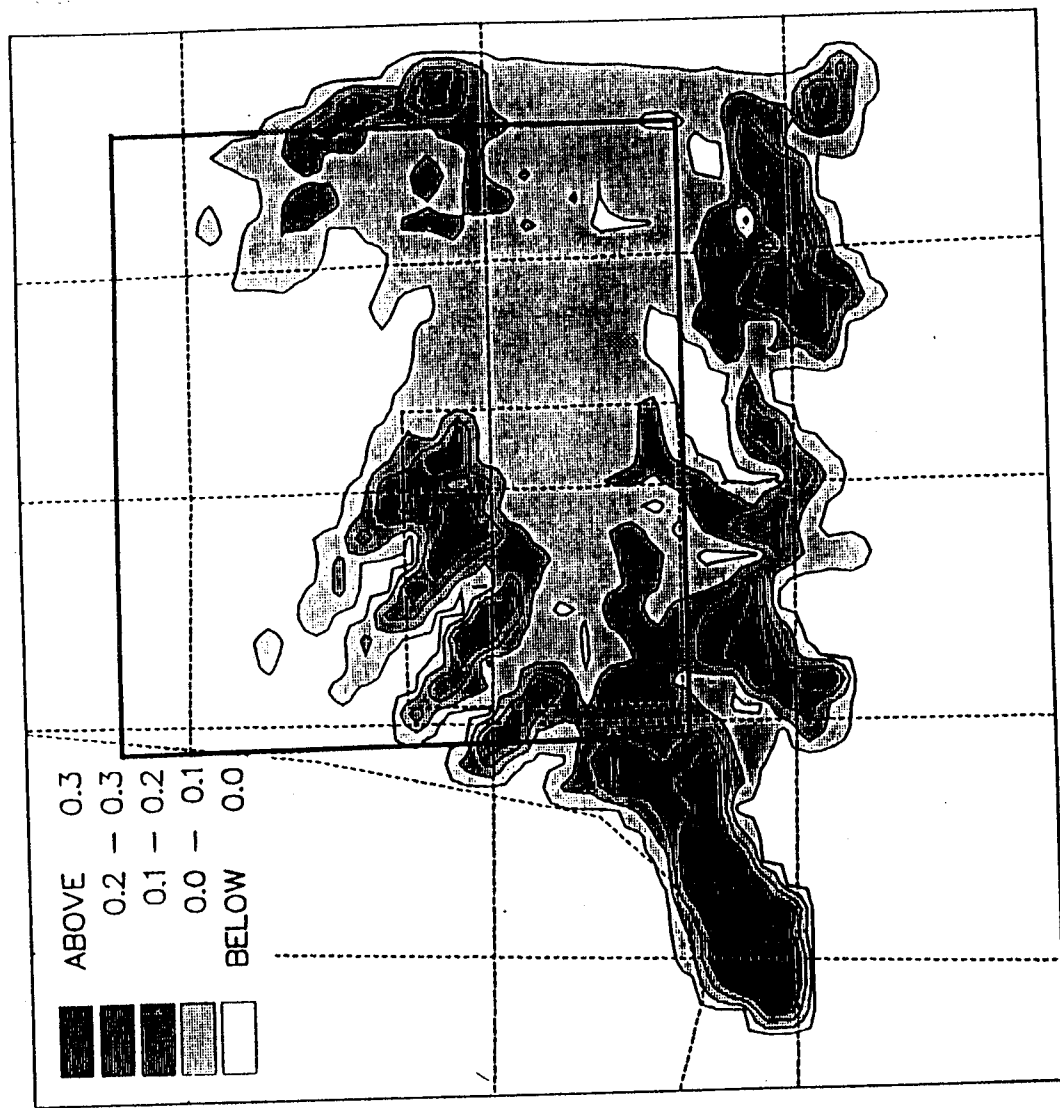
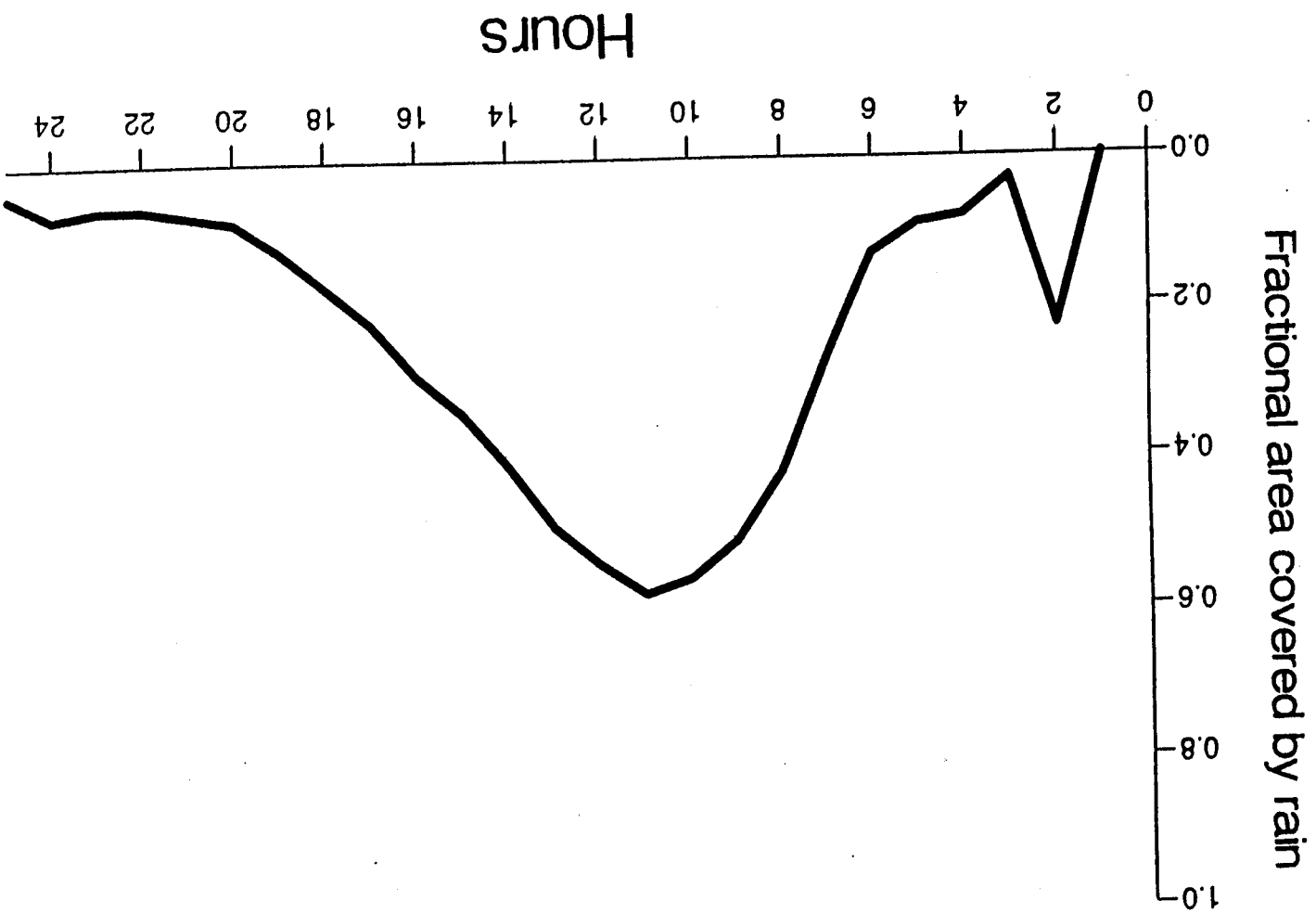


Figure 7

Figure 8



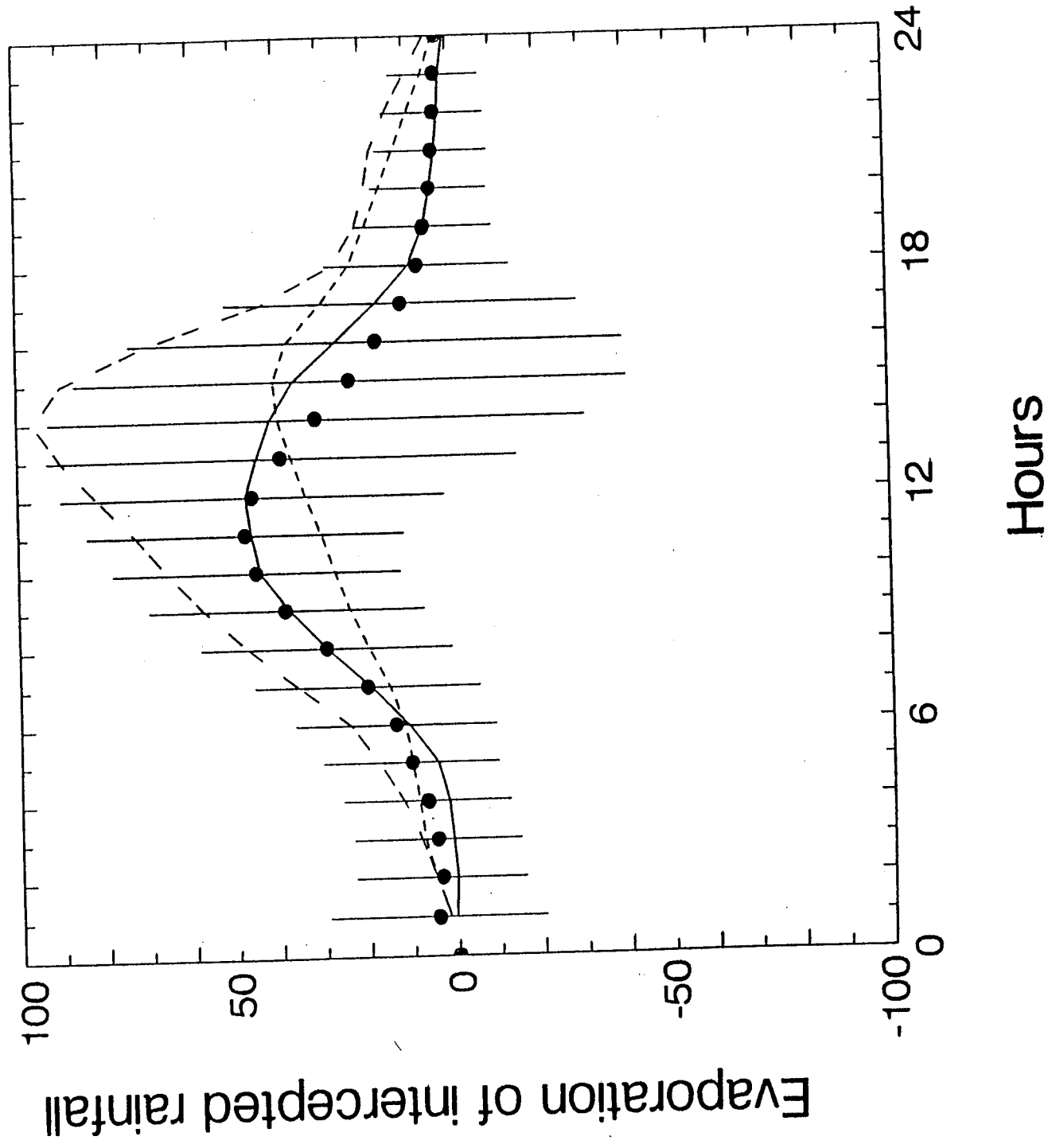


Figure 9

Mesoscale Fluxes Over Heterogeneous Flat Landscapes For Use In Larger Scale Models

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Abstract

We have developed a framework to parameterize mesoscale fluxes of heat, moisture, and momentum in a cloud-free environment which result from heterogeneous heating of flat land surfaces. The importance and parameterizability of these mesoscale fluxes is demonstrated using the mathematical concept of predictability. This methodology is used to estimate the relative importance of mesoscale, as contrasted with turbulent fluxes, in the Konza Prairie of Kansas during the FIFE field experiment.

1. Introduction

During the last decade or so, procedures to represent landscape heterogeneity within a larger-scale area, such as a general circulation model (GCM) grid cell, have focused on a summation of surface fluxes within that area, proportionally weighted by the fluxes from each land surface type. Examples of this approach are reported in Avissar and Pielke (1989), Li and Avissar (1994), Kosta and Suarez (1992), Bonan et al. (1993), Henderson-Sellers et al. (1993), Miller (1994), and Arritt and Clark (1994).

There is convincing modeling evidence, however, that mesoscale¹ fluxes that result from landscape heterogeneity in flat terrain can often be as large as and larger than turbulent fluxes, as well as have a different vertical distribution than the turbulent fluxes averaged over the scale of a GCM grid such as illustrated in Figure 1. Examples of studies that document the modification of atmospheric boundary layer structure, and/or the development of mesoscale flow due to land surface inhomogeneity are reported in Pielke et al., 1991, 1993a; Dalu et al., 1991; Cotton and Pielke, 1992; Segal and Arritt, 1992; Avissar and Pielke, 1989; Manqian and Jinjun, 1993; Raupach, 1991; Guo and Schuepp, 1994; Avissar and Chen, 1993 and André et al., 1989. A number of these papers are summarized in Cotton and Pielke (1992). These mesoscale fluxes have substantial coherent structure such that they are predictable features (i.e., Zeng and Pielke, 1993a; Zeng, 1992). Observations have documented the importance of heterogeneous landscape in influencing boundary layer structures and mesoscale fluxes (Mahrt, 1994; Mahrt and Ek, 1993; Mahrt et al., 1994a, b; Doran et al., 1992; Smith et al., 1992; Beljaars and Holtslag, 1991; Segal et al., 1988, 1989; and Balling, 1988) and in effecting such related properties as soil water infiltration (Wood et al. 1992).

¹We define mesoscale using the definition presented in Pielke (1984). Mesoscale systems can be accurately described by the hydrostatic equations of motion, while the turbulent motions are nonhydrostatic features. Mesoscale features differ from larger scale systems in that the horizontal winds are substantially out of gradient wind balance even above the planetary boundary layer.

Therefore, it should be possible to parameterize the influence of the mesoscale fluxes on larger scale atmospheric structures. This is the purpose of this paper. Clausen (1991), for example, has discussed how land surface variability influences spatially averaged fluxes and spatially averaged vertical gradients, although his analysis is limited to the situation where the boundary layer becomes homogeneous above a "blending height" (Wierenga, 1986), which is situated within the surface layer. Mahrt (1987) and Mason (1988) also have papers that are concerned with this topic. Lynn et al. (1994) are also developing a new procedure to parameterize mesoscale fluxes resulting from landscape patches.

Field programs which focus on land surface-atmosphere interactions such as FIFE (Sellers et al., 1992a, and Betts and Beljaars, 1993) in Kansas, U.S.A.; BOREAS (see Experiment Plan, 1993) in Saskatchewan and Manitoba, Canada; LOTREX (Schädler et al., 1990) in Hildesheimer Börde, Germany; EFEDA (Bolle et al., 1993) in Spain; HAPEX-MOBILHY (André et al., 1989; Noilhan et al., 1991) in France; HAPEX-NIGER92 (e.g., Gash et al., 1991); SEBEX (Wallace et al., 1991) in the Sahel of Africa and in the Amazon region (e.g., Shuttleworth, 1985; Wright et al., 1992; Gash and Shuttleworth, 1991) can provide an observational data base to contrast with the model results.

2. Methodology

Linear studies can provide us with some insight as to when mesoscale circulations are likely to generate vertical fluxes of heat, moisture, and momentum on the same order as turbulent fluxes. Dalu and Pielke (1993) report on one such modeling effort where linear response of the atmosphere to heat patches of various sizes in flat terrain are examined. Based on that analysis and for the case of zero large scale flow, horizontal fluxes can homogenize the atmosphere a short distance above the surface (such that horizontal gradients in boundary layer structure do not develop) when:

$$K_z \left(\frac{m\pi}{R_0} \right)^2 T F_1(t_m) = \text{Order } (1), \quad (1)$$

where R_0 is the Rossby radius of deformation, T is proportional to the inertial period and a period related to deceleration of horizontal flow, $m = 2R_0/L_z$, and $F_1(t_m)$ is the time dependent evolution of the mesoscale circulation. K_z is the horizontal exchange coefficient. L_z is the horizontal spatial scale of the heat patches. Claussen (1991) defines the height at which the atmosphere homogenizes (i.e., the blending height) as the level with the same scale as the vertical diffusion length scale.

The product of variables in Eq. (1) is of order unity, as discussed in Dalu and Pielke (1993), when:

$$m = \frac{1}{\pi} \left(\frac{N_0 h_0 R_0}{F_1(t_m) K_z} \right)^{\frac{1}{2}},$$

where $N_0 = [(g/\theta)\partial\theta/\partial z]^{\frac{1}{2}}$ (i.e., the Brunt-Väisälä frequency) and h_0 is the maximum depth attained of the planetary boundary layer. Using these relations, at high wave-numbers (i.e., large m), for example with $m = 100$ ($L_z = 2$ km), the mesoscale fluxes from this patch size would be important all day for $K_z = 10 \text{ m}^2 \text{ s}^{-1}$ (for this example, $R_0 = 100$ km, $h_0 = 1.5$ km, $T = 10^4$ s, and $N_0 = 10^{-2} \text{ s}^{-1}$ were defined). If $K_z = 100 \text{ m}^2 \text{ s}^{-1}$, however, only patches having a wavenumber less than or equal to 30 ($L_z = 7$ km) would produce significant mesoscale fluxes throughout the day. With this large value of K_z , the $L_z = 2$ km patches would have significant fluxes relative to the turbulent fluxes only for a very short time in the morning, if the diurnal heating of the patches was the dominant surface forcing mechanism. This analytic model indicates that mesoscale fluxes (and thus, the need to parameterize them) become insignificant when the horizontal scale of the flow is on the same order as the vertical scale (i.e., $R_0/m \cong h_0$). Avissar and Pielke (1989) assumed this result, but without theoretical justification, in developing a mosaic one-dimensional representation of land surface effects.

Using mixed layer scaling (e.g., see Garratt, 1992), the horizontal exchange coefficient can be estimated as:

$$K_z \cong l\sigma_H$$

where for a mixed layer $\sigma_H = u_* (12 + 0.5h_0/|L|)^{\frac{1}{2}}$ and $l = 1.5 h_0$ (e.g., see Pielke, 1984, pgs. 177-178) can be used. The parameter u_* is the friction velocity and L is the Monin length. From this analysis, as the heat patches become smaller, their influence is more easily mixed horizontally, as a function of height above the surface (as an example, with $u_* = 0.5 \text{ m s}^{-1}$, $h_0 = 1.5 \text{ km}$, $|L_z| = 10 \text{ m}$, yields $K_z = 8.55 \times 10^3 \text{ m}^2 \text{ s}^{-1}$). This intensity of horizontal mixing also depends on the thermodynamic stability in the lower atmosphere, in which more vigorous convective heating would result in stronger horizontal mixing.

The introduction of large scale wind flow modifies the importance of mesoscale fluxes, as shown by Pielke et al. (1993b), with stronger mesoscale fluxes where the large scale flow is opposite in direction and about the same magnitude of speed as the resultant mesoscale induced horizontal wind flow. Otherwise, the presence of a large scale wind reduces the importance of mesoscale fluxes for the same size surface heat patch scale. This large scale wind dilutes the influence of the heat patches when they are sufficiently small by advecting warm air over colder surfaces, and colder air over the warmer patches, thereby promoting homogenization above the surface as a result of this advection and destabilization of the atmosphere over the warmer surfaces. This effect was demonstrated in a set of large eddy model simulations over flat terrain reported in Hadfield et al. (1991, 1992).

To determine whether we can parameterize mesoscale and turbulent fluxes associated with landscape patchiness, we first have explored the predictability of these flows. The predictability measure that we apply is the signal-over-noise ratio, r . The signal is defined to be the domain-averaged root mean square (rms) difference of a variable at time t and at the initial time, and the noise is defined to be domain-averaged rms difference of a variable

in the control simulation and in the perturbed simulation at time t . The domain average is obtained from a vertical and horizontal summation of each grid point in the domain. In order to estimate the overall predictability during the day, we have also computed the mean value $\bar{r}$, of the signal-over-noise ratio $r(t)$ from $t = 3$ to 11 hours. When $\bar{r}(a) \geq 1$ the variable a is defined to be predictable during the day, otherwise the variable a is unpredictable. Figures 2-4 show ($\bar{r}$) of θ , u , and w as a function of the synoptic wind U and the patch size L_x , respectively. It is seen that $\bar{r}(\theta)$ is always larger than unity. For most part of the parameter space, $\bar{r}(\theta)$ is about 10. $\bar{r}(u)$ is greater than unity except when both U and L_x are small. For most of the parameter space, $\bar{r}(w)$ is greater than unity.

In Figs. 2-4, when L_x is small, $\bar{r}$ is relatively small because surface differential heating is small. In Figs. 3 and 4 (for u and w), as L_x increases, $\bar{r}$ decreases. The reason is that when L_x is large, except near the coastal area, the domain is dominated by boundary layer turbulence, just like the homogeneous case. This has been discussed in Zeng and Pielke (1993a). Further discussions on the predictability of surface-induced flow are given elsewhere (Zeng and Pielke, 1993b; 1994a).

Since there exists predictability for a wide range of values of synoptic wind flow and patch size, this indicates that we should be able to develop a parameterization of mesoscale and turbulent fluxes over this parameter space.

Analogous to mixed layer boundary layer scaling and similarity scaling, we hypothesize that there exists non-dimensional scaling parameters that can be used to evaluate the importance of mesoscale fluxes over heterogeneous, flat terrain.

The assumed non-dimensional parameters to be functionally related to the mesoscale fluxes are:

$$\begin{aligned} R_b &= (gz_i \delta \theta) / (\theta_* U^2) \\ z_i / L_x \\ \lambda_1 &= U z_i / L_x w_* \\ \lambda_2 &= U^2 / c_p \Delta \theta \end{aligned}$$

U is the large scale advection wind estimated as the synoptic flow at the boundary layer height z_i , while $\Delta\theta$ is the maximum horizontal surface temperature difference at the roughness height z_0 , between different patches. L_x is the dominant horizontal patch size and $\delta\theta$ is the potential temperature difference between the depth z_i , and the surface. The potential temperature at z_0 is θ_* , and $w_* = (gQ_* z_i / \theta_*)^{1/3}$ where Q_* is the turbulent surface sensible heat flux. The parameter, w_* , is the convective velocity scale. Each of these parameters can be estimated from the resolvable variables in larger scale models, except $\Delta\theta$ and L_x . These two parameters, however, could be obtained using a mosaic approach (e.g., Avissar and Verstraete, 1990; Avissar and Pielke, 1989; and Miller, 1994) where vegetation, soil, and water coverage information are used to obtain such information as leaf area index, soil moisture, etc. within sub-areas of a larger scale model grid cell. R_i provides a measure of the intensity of convective turbulent generation while z_i/L_x is a measure of the depth, z_i , into the atmosphere in which this turbulence would extend relative to the dominant horizontal patch size, L_x . Horizontal gradients in turbulent sensible heating, and the depth, z_i , to which this heating extends are what generates mesoscale circulations. λ_1 yields a measure of the relative importance of horizontal advection to convective vertical mixing, while λ_2 is a ratio related to horizontal advection and the horizontal pressure gradient force, as introduced in Pielke (1984, Chapter 3).

Besides these four independent parameters, we have also suggested another four parameters and they are:

$$\begin{aligned}\lambda_3 &= \lambda_2 z_i / (\lambda_1 L_x) = U w_* / (c_p \Delta\theta) \\ \lambda_4 &= \lambda_2 R_i = g z_i \delta\theta / (\theta_* c_p \Delta\theta) \\ \lambda_5 &= \lambda_2 L_x / z_i = L_x U^2 / (z_i c_p \Delta\theta) \\ \lambda_6 &= \lambda_1 L_x / z_i = U / w_*\end{aligned}$$

We expect that mesoscale fluxes would vary in a nonlinear fashion when these parameters vary, but would be a well-defined function of these effects. For example, mesoscale effects would be expected to dominate for intermediate values of L_x (Xian and Pielke, 1991; Avissar

and Chen, 1993). For very small values of L_z , the atmosphere is quickly homogenized immediately above the surface such that mesoscale circulations do not develop (e.g., Hadfield et al., 1991; 1992). For very large values of L_z , mesoscale circulations would develop only near the patch edge with the interiors dominated by turbulent fluxes. This paper places quantitative limits on the influence of these parameters on the flux.

In order to explore the parameter space of the above parameters, in order to assess the value of this parameterization approach, and to study the relationship between mesoscale, turbulent, and total fluxes, we have performed extensive two-dimensional numerical models by means of the Colorado State University-Regional Atmospheric Modeling System (CSU-RAMS; Pielke et al., 1992). The nonhydrostatic equations are used with moisture taken as a passive tracer. Radiation, subgrid scale turbulence, and soil model parameterizations are included. The horizontal grid intervals are $\Delta x = 2$ km, and periodic lateral boundary conditions are used. A more detailed model description has been given in Zeng and Pielke (1993a). The surface consists of alternate strips of water and land with strip sizes of $L = 10, 20, \dots, 100$, and 150 km. The synoptic wind is $U = 3, 5, 7$, and 9 m s^{-1} . When U is smaller than 3 m s^{-1} , some of the above parameters are either too small or too large, and the above parameters need to be modified; e.g., replacing U by $(U + w_*)$. This issue will be addressed in our future work.

Figure 5 illustrates how the values of the eight parameters vary as a function of the horizontal scale (L) of surface heat patches and the synoptic wind U . Relatively smooth curves are observed in Fig. 5, which is a necessary condition in order to properly parameterize different fluxes as a function of these parameters. Figure 5 also represents the part of the parameter space that we have explored.

3. Parameterization Form

While there still needs to be additional model runs to explore the entire spectrum of meteorological conditions, there is sufficient robustness to the model runs to present the functional form of the algorithm for the large-scale averaged mesoscale, turbulent, and total fluxes.

These fluxes are strongly dependent upon the dimensionless height z/z_i . Schematically, our parameterization equation for any dimensionless flux F can be written as

$$F - f_1(z/z_i) = g\left(\frac{z_i}{L}, R_b, \lambda_1, \dots, \lambda_6\right) f_2(z/z_i) \quad (2)$$

In our study, the function f_1 is a quadratic or cubic polynomial and the function f_2 is a linear or quadratic polynomial (or this polynomial multiplied by $f_1(z/z_i)$). The function g is a linear or quadratic polynomial of each parameter.

There are two conflicting requirements on a parameterization scheme: the scheme should use a functional form as simple as possible, and should use as few parameters as possible; and the scheme should be as accurate as possible. To balance these two requirements, we propose the following parameterization equations for domain-averaged (i.e., large-scale model grid-box-averaged), mesoscale (with the superscript " ' "), turbulent (with the superscript " "), and total fluxes of sensible heat $< w\theta >$, latent heat $< wq >$, and momentum $< wv >$ in the boundary layer. These equations have been tested against model-generated data at time $t = 7, 7.5$, and 8 hours after sunrise based on statistical techniques which are discussed in Zeng and Pielke (1994b).

For *mesoscale fluxes*, which are defined in this paper as the resolvable flux with respect to model grids with $\Delta x = 2$ km, and *small-scale (or turbulent) fluxes*, which are defined as the subgrid parameterized fluxes in this model grid, the general formula is

$$\begin{aligned} \frac{F-f_1\left(\frac{z}{z_i}\right)}{f_1\left(\frac{z}{z_i}\right)} &= [a_1 + a_2p_1 + a_3p_1^2 + a_4p_2 + a_5p_2^2 + a_6p_3 + a_7p_3^2] \\ &+ [b_1 + b_2p_1 + b_3p_1^2 + b_4p_2 + b_5p_2^2 + b_6p_3 + b_7p_3^2] \frac{z}{z_i} \end{aligned} \quad (3)$$

where F is a flux scaled by its total (or turbulent) surface value, p_i ($i = 1, 2$, and 3) represent three of the eight parameters (R_b , z_i/L_z , λ_1 , λ_2 , λ_3 , λ_4 , λ_5 , λ_6) and

$$f_1\left(\frac{z}{z_i}\right) = c_1 + c_2\left(\frac{z}{z_i}\right) + c_3\left(\frac{z}{z_i}\right)^2 + c_4\left(\frac{z}{z_i}\right)^3 \quad (4)$$

The values of coefficients a_i , b_i , c_i , and parameters p_i for mesoscale and turbulent fluxes are given in Tables 1-4.

For the *total sensible heat flux* scaled by its surface value, the formula is

$$\begin{aligned} \frac{F-f_1\left(\frac{z}{z_i}\right)}{f_1\left(\frac{z}{z_i}\right)} &= [a_1 + a_2\lambda_2 + a_3\lambda_3 + a_4\lambda_4 + a_5\lambda_6 + a_6R_b] \\ &+ [b_1 + b_2\lambda_2 + b_3\lambda_3 + b_4\lambda_4 + b_5\lambda_6 + b_6R_b] \frac{z}{z_i} \end{aligned} \quad (5)$$

where

$$f_1\left(\frac{z}{z_i}\right) = c_1 + c_2\left(\frac{z}{z_i}\right) + c_3\left(\frac{z}{z_i}\right)^2 \quad (6)$$

The values of coefficients a_i , b_i , and c_i are given in Table 5.

For the *total moisture flux* scaled by its surface value, the formula is

$$\begin{aligned} \frac{F-f_1\left(\frac{z}{z_i}\right)}{f_1\left(\frac{z}{z_i}\right)} &= \left[a_1 + a_2\lambda_1 + a_3\lambda_1^2 + a_4\lambda_4 + a_5\lambda_4^2 + a_6\lambda_6 + a_7\lambda_6^2 + a_8\left(\frac{z_i}{L_z}\right)^2 + a_9\left(\frac{z_i}{L_z}\right)^2 \right] \\ &+ \left[b_1 + b_2\lambda_1 + b_3\lambda_1^2 + b_4\lambda_4 + b_5\lambda_4^2 + b_6\lambda_6 + b_7\lambda_6^2 + b_8\left(\frac{z_i}{L_z}\right) + b_9\left(\frac{z_i}{L_z}\right)^2 \right] \frac{z}{z_i} \end{aligned} \quad (7)$$

where $f_1 \left(\frac{z}{z_i} \right)$ is a quadratic polynomial as expressed in Eq. (6). The values of the coefficients a_i , b_i , and c_i in Eqs. (6) and (7) are given in Table 6.

For the *total momentum flux* scaled by its surface value F , the formula is

$$\begin{aligned} F - f_1 \left(\frac{z}{z_i} \right) &= \left[a_1 + a_2 \lambda_4 + a_3 \lambda_6 + a_4 \left(\frac{z_i}{L_s} \right) \right] \\ &+ \left[b_1 + b_2 \lambda_4 + b_3 \lambda_6 + b_4 \left(\frac{z_i}{L_s} \right) \right] \frac{z}{z_i} \\ &+ \left[c_1 + c_2 \lambda_4 + c_3 \lambda_6 + c_4 \left(\frac{z_i}{L_s} \right) \right] \left(\frac{z}{z_i} \right)^2 \end{aligned} \quad (8)$$

where

$$f_1 \left(\frac{z}{z_i} \right) = d_1 + d_2 \left(\frac{z}{z_i} \right) + d_3 \left(\frac{z}{z_i} \right)^2 + d_4 \left(\frac{z}{z_i} \right)^3 \quad (9)$$

The coefficients a_i , b_i , c_i , and d_i are given in Table 7.

To demonstrate an application of the parameterization scheme, we use Eqs. (3) and (4) to assess the relative importance of mesoscale and turbulent fluxes for the FIFE study area in Kansas. FIFE field program results and an overview of the program are presented in Sellers et al. (1992a). Using results presented in Smith et al. (1994) and Betts et al. (1992), the parameters required for use in Eqs. (3) and (4) were estimated as: $L = 10\text{--}20$ km, $w_* = 1.6 \text{ m s}^{-1}$, $\delta\theta = -3^\circ\text{C}$, $\Delta\theta = 5^\circ\text{C}$, $z_i = 1$ km, and $U = 4 \text{ m s}^{-1}$. Values of these parameters, of course, vary from day to day but these estimates will provide insight as to the relative importance of mesoscale fluxes in that study area.

Using $L = 10$ km and 20 km, the ratio of mesoscale to turbulent fluxes of heat, moisture, and momentum as a function of height is presented in Tables 8 and 9. It is seen that near the surface the turbulent fluxes are dominant; however, in the upper portion of the boundary layer the mesoscale fluxes (especially those of sensible heat and momentum) are dominant. This conclusion is consistent with the interpretation of Smith et al. (1994) and others who participated in the FIFE study, in that mesoscale fluxes did occur. This is also

consistent with the estimate of FIFE-area surface fluxes obtained from FIFE-area averaged variables which are essentially identical to the linear averages of individual point specific fluxes, as reported in Sellers et al. (1992b), because the total fluxes are always dominated by turbulence, and the mesoscale ascent is small near the surface.

4. Conclusions

A procedure to represent mesoscale and subgrid scale fluxes in a cloud-free environment over flat heterogeneous landscapes is introduced. New non-dimensional parameters were presented with which a parameterization of this effect for use in larger-scale models was constructed. While additional parameters, such as the Coriolis term f need to be considered, and wind speeds less than 3 m s^{-1} need to be included, the framework presented in this paper has an apparent universal form which should be used for further study, and validation against observations. Also the generality of the empirical coefficients used to construct the parameterizations needs to be assessed.

One of the conclusions is that one can ignore mesoscale fluxes when they are a small fraction of the turbulence fluxes. For these situations, the one-dimensional parameterizations introduced by Avissar and Pielke (1989), Sellers et al. (1992b), or Miller (1994) could be used.

5. Acknowledgments

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← 6. Discussion

Question: (Fiedler) I would like to refer to the statement that topographical effects are weaker than surface albedo. Does this statement include the effects (observable and modeled) of triggering convection by lee waves or Fohn effects with extreme variability of precipitation and surface heating in different elevations?

Response: Our results related to the relative importance of albedo versus terrain horizontal variations were for zero large scale flow as summarized in our paper (Pielke, R.A., J.H. Rodriguez, J.L. Eastman, R.L. Walko and R.A. Stocker, 1993: Influence of albedo variability in complex terrain on mesoscale systems. *J. Climate*, 6, 1798-1806). We limited our model runs to zero synoptic flow in order to isolate the significance of terrain acting as an elevated heat source, from albedo heterogeneity. For substantial non-zero large scale wind, the orographic effect of terrain, of course, becomes dominant.

Question: (Schaut) Although you have stated that albedo effects can be as large as the effects of topography, the topographical effects dominate in your northern Arizona data. I believe you stated, in answer to another's question, that the albedo effects dominate over a relatively flat surface only when there are no horizontal winds. I also believe you stated that horizontal winds tend to reduce the effects of albedo differences. Given that mountains tend to produce winds, could this be at least part of the reason the winds in northern Arizona appear to be dominated by topography?

Response: As I responded to Professor Fiedler, our study with respect to albedo/terrain effects in northern Arizona were for zero large scale flow. The actual observed flow in that region, of course, is strongly influenced by terrain since the synoptic winds are often not weak. Also, even when the large scale flow is near zero, the heating of the elevated terrain is on the same order as the albedo heating effect, so that the terrain configuration will have a substantial influence on the local wind flow. The feedback over time of vegetation and snow cover, and precipitation patterns which evolve in response to the albedo and terrain configurations, should be important and represents an important area of research.

Question: (Choudhury) How important is it to consider differences of roughness heights for heat, mass, and momentum in mesoscale land-atmosphere interaction? How will spatial heterogeneity of these differences in the roughness height impact the interaction?

Response: The importance of different roughness heights for heat, moisture, and momentum with respect to mesoscale land-atmosphere interactions needs to be explored. This procedure to represent surface fluxes is being added to RAMS.

Question: (Schaaake) This is a question concerning your work to develop parameterizations of the effects of sub-grid variability of heat patches that have a representative size of, say, m in an atmospheric model having a grid mesh length of Δx . Does the parameterization depend on the ratio $\Delta x/m$? Is there a value of $\Delta x/m$ above which the parameterization would not depend on Δx ? Does the vertical resolution influence the parameterization?

Response: The influence of grid resolution on the procedure to parameterize the effects of the sub-grid scale variability of heat patches depends on the spatial scale of the heat patches and the grid size. For small enough patches (i.e. large m) our work (as well as that of others) indicates that we can fractionally weight the surface fluxes over whatever grid area is used. When mesoscale fluxes become important, however, care must be taken not to "double count" these fluxes since for intermediate grid sizes (e.g. $\Delta x = 10\text{-}20\text{ km}$), part of the mesoscale flux could be subgrid scale while a component is explicitly resolved by the model. A similar concern has been recognized with respect to cumulus parameterization schemes.

Vertical resolution will influence the parameterization in that the structure with respect to height of both the turbulent and mesoscale fluxes must be adequately represented in order, for example, to accurately simulate the heating rates in the vertical.

Question: (Schulze) Over what distance do you see an advection effect of adjacent vegetation (or ocean) on actual fluxes of a given plot?

Response: The distance over adjacent vegetation (or ocean) in which the upstream surface conditions are felt depends on the thermodynamic stability and adjective wind speed near the surface over the vegetation (or ocean). If the underlying surface is thermodynamically unstable, the adjustment to the new surface will be comparatively rapid. If the surface is relatively cold, however, such that it is thermodynamically stable, the upstream conditions can propagate quite a distance over the new surface. I discuss this question in Pielke (1984, Chapter 7, *Mesoscale Meteorological Modeling*, Academic Press).

Question: (Dolman) Are you using in your linear analysis a background wind speed or do you assume this is zero? How will this affect your conclusion about the small influences of formulated advection?

Response: We have analyzed cases with both zero and non-zero background winds in our analytic model. Results are reported in: Dalu and Pielke (1993): Vertical heat fluxes generated by mesoscale atmospheric flow induced by thermal inhomogeneities in the PBL. *J. Atmos. Sci.*, **50**, 919-926; Dalu, G.A., M. Baldi, R.A. Pielke, T.J. Lee, and M. Colacino, (1991): Mesoscale vertical velocities generated by stress changes in the boundary layer: Linear theory, *Ann. Geophys.*, **9**, 648-653; and in Pielke, R.A., G. Dalu, T.J. Lee and H. Rodriguez, J. Eastman, and T.G.F. Kittel, (1993): Mesoscale parameterization

of heat fluxes due to landscape variability for use in general circulation models. *Proc. Joint Scientific Assembly of the International Association of Meteorology and Atmospheric Physics and the International Association of Hydrological Sciences*, July 11-13, 1993, Yokohama, Japan.

Question: (Raupach) Regarding simulations of thunderstorm development in Georgia, and effect of land use change and timing of onset of storm activity: Can we distinguish between effects of mean change in land properties (e.g. albedo, soil moisture, latent heat flux) and variability of land properties? By variability I mean both magnitude (e.g. standard deviation) and length scale.

Response: In our north Georgia simulation, we have thus far only explored the current and a derived potential landscape condition. Thus, both the "mean" conditions and the spatial variability of land properties would be different between the experiments. It would be valuable from a scientific perspective to separate these two effects, in order to assess their relative importance, and we plan to do that in response to Dr. Raupach's suggestion.

Question: (Becker) Pielke demonstrated that there are land surface heterogeneities of a certain length scale (in this case islands of the size of 10 to about 100 km) which influence in a predictive way the circulation pattern and thus the regional climate. This may open the possibility to implement such "fixed" structures into larger scale atmospheric models and improve their capability to represent this type of land surface heterogeneity.

Response: I agree with Dr. Becker that predictive skill in weather forecasting could be improved by incorporating our knowledge of how the atmosphere responds to fixed geographic features into larger scale models.

Comment: (Schulze) I support this view that effects of land use are stronger than effects of doubling CO₂. The physiological reason for this is the effect of plot nutrition on assimilation and surface conductance. Natural vegetation has a Nitrogen concentration of about 1% while agricultural crops have more than 3%; this drives surface conductance more than the content of sand in soil. This is important if we compare non-fertilized areas (e.g. Australia) with fertilized areas such as Europe (reference: Annual Res. Ecol. Systems, 1995, in press).

Question: (Kabat) The speaker stressed a need to couple general circulation models (and mesoscale models) with models of land surface dynamics (hydrology, vegetation) and to use the coupled models for impact effect studies.

Response: We are working on coupling RAMS to ecological and hydrological models. Indeed, there has been considerable recent interest in such coupled models which are essential if we are to perform more realistic climate change scenarios.

Comment: (Kabat) Up to this point there is neither a general circulation model nor a mesoscale circulation model which is fully coupled to a model of vegetation dynamics and growth. There is not even a coupled model for a stable ecosystem (no change in species) on the time scale of more than a growing season. So it seems to be a long way to go before this can be confirmed.

7
6. References

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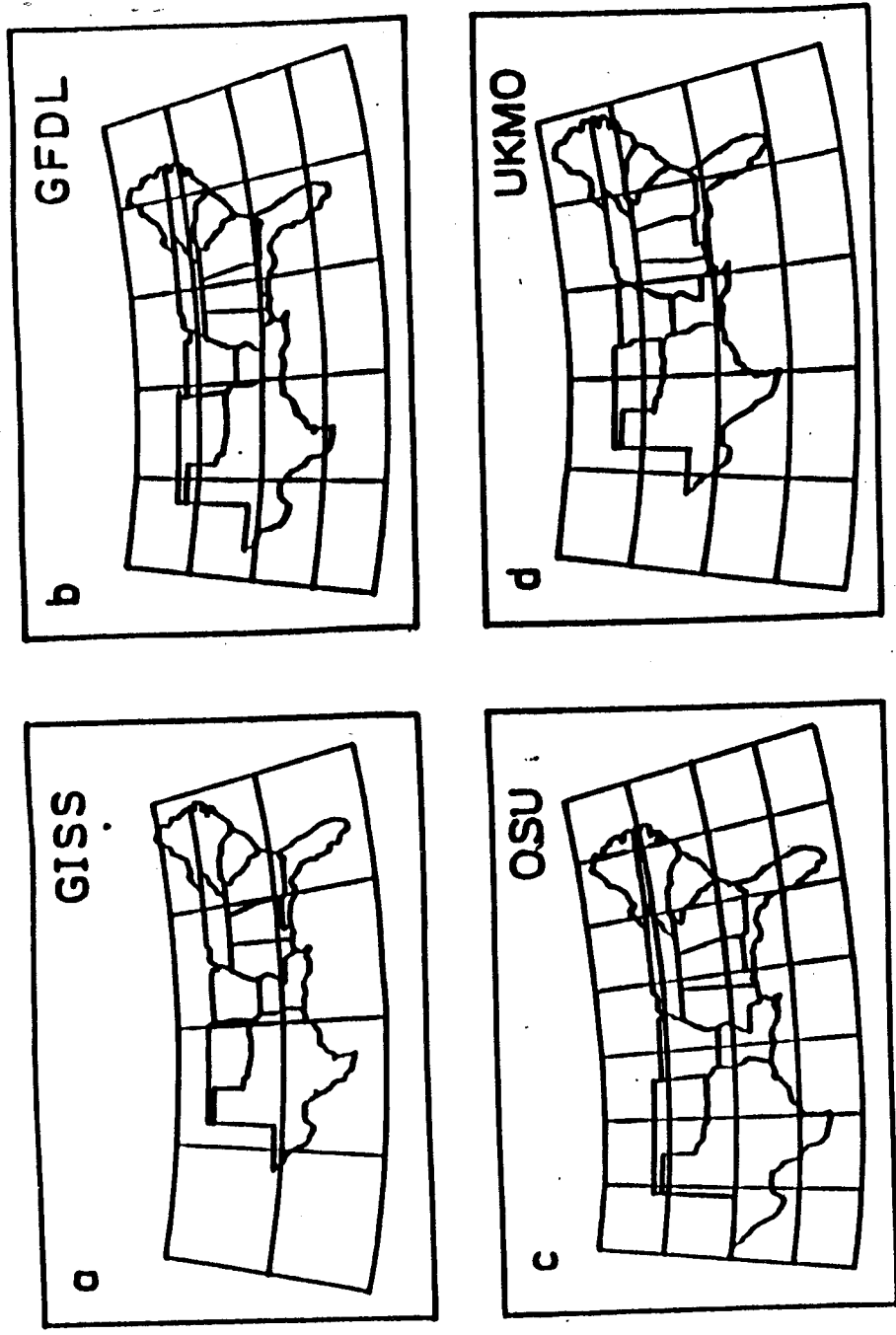
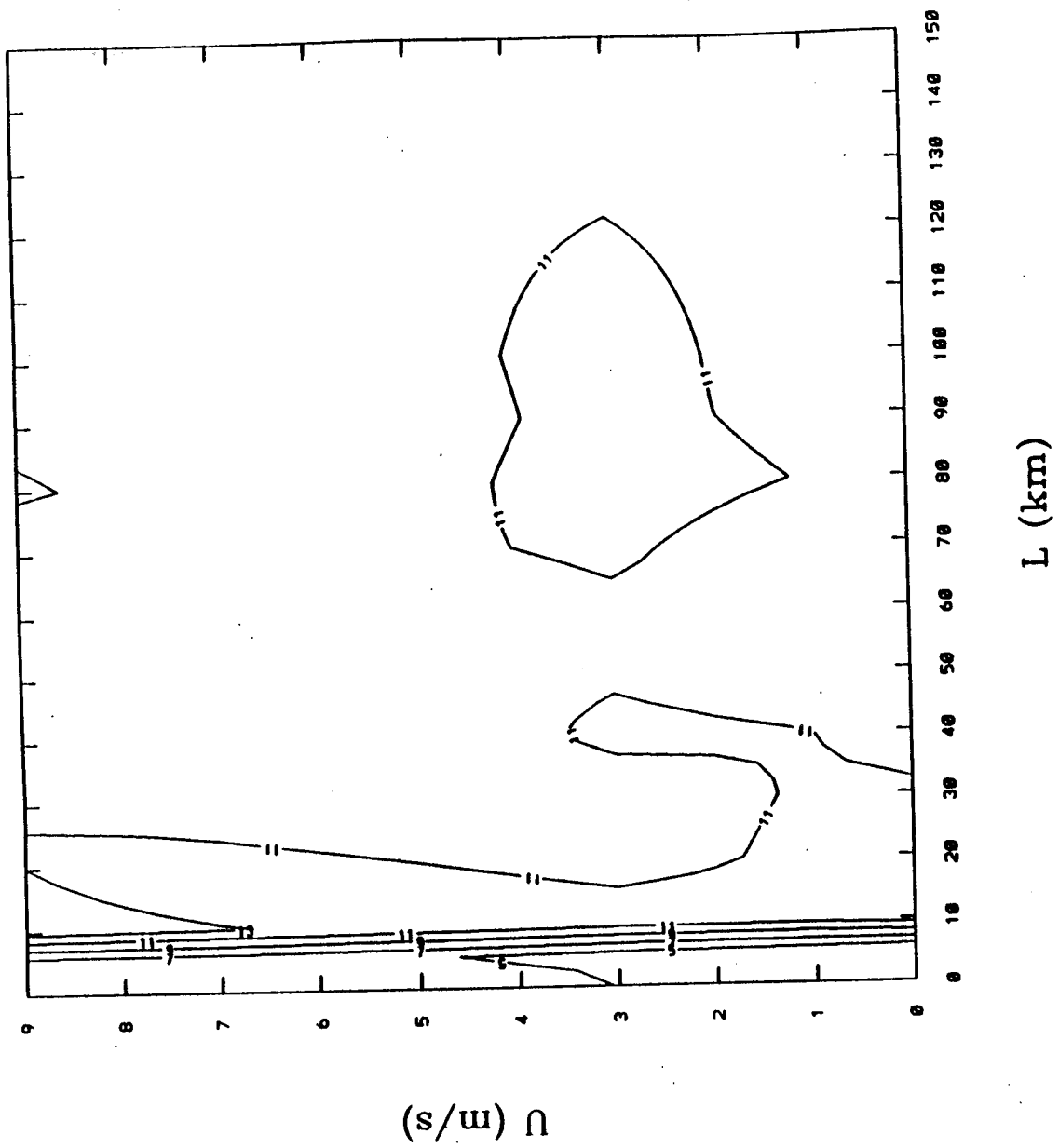
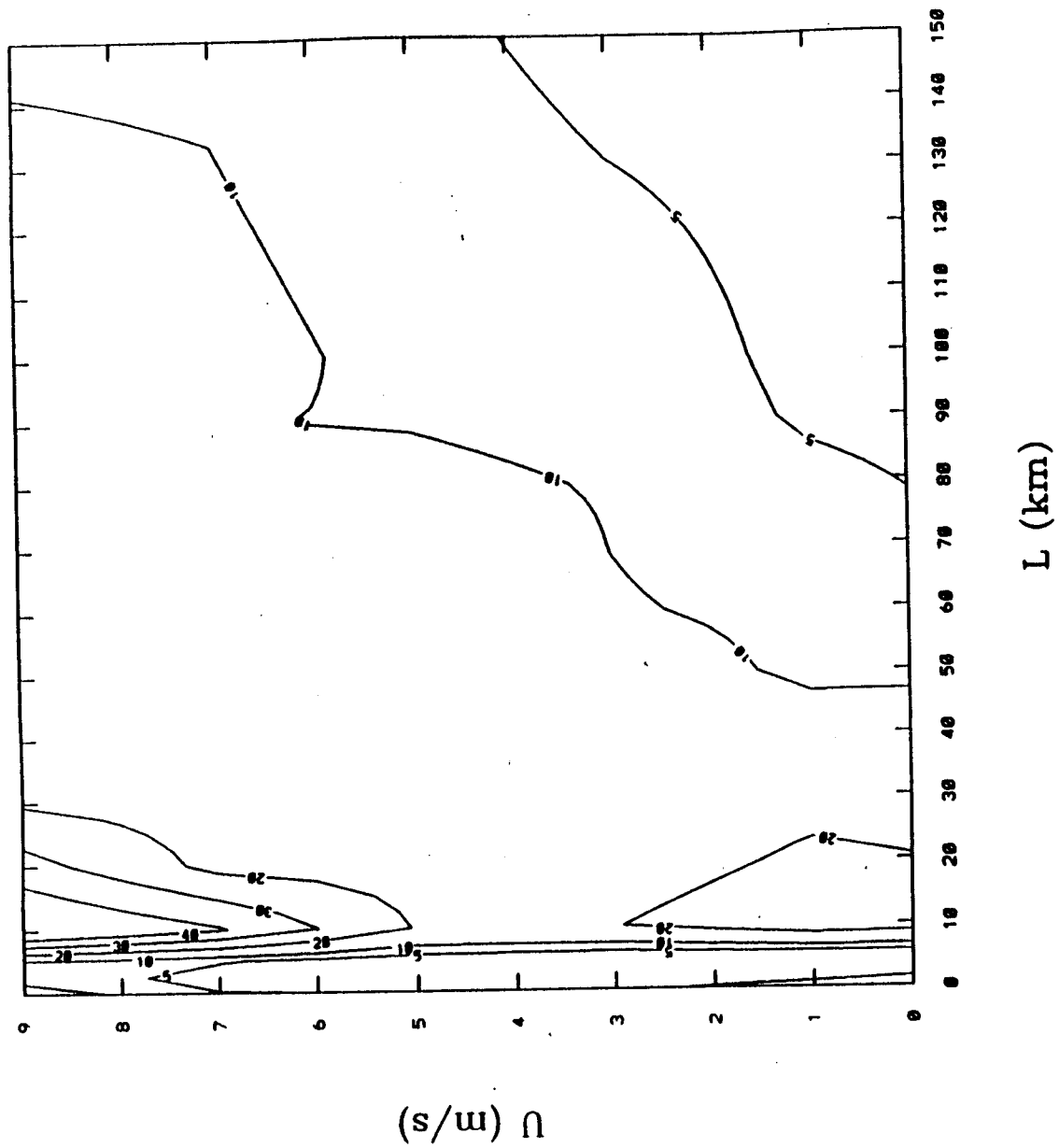


Figure 1: Horizontal resolution of four general circulation models (GCMs) across the southern U.S.: (a) GISS, (b) GFDL, (c) OSU, and (d) UKMO (from Cooter, E.J., B.K. Eder, S.K. LeDuc, and L. Truppi, 1993: General circulation model output for forest climate change research and applications. Southeastern Experiment Station Technical Bulletin, in press.)





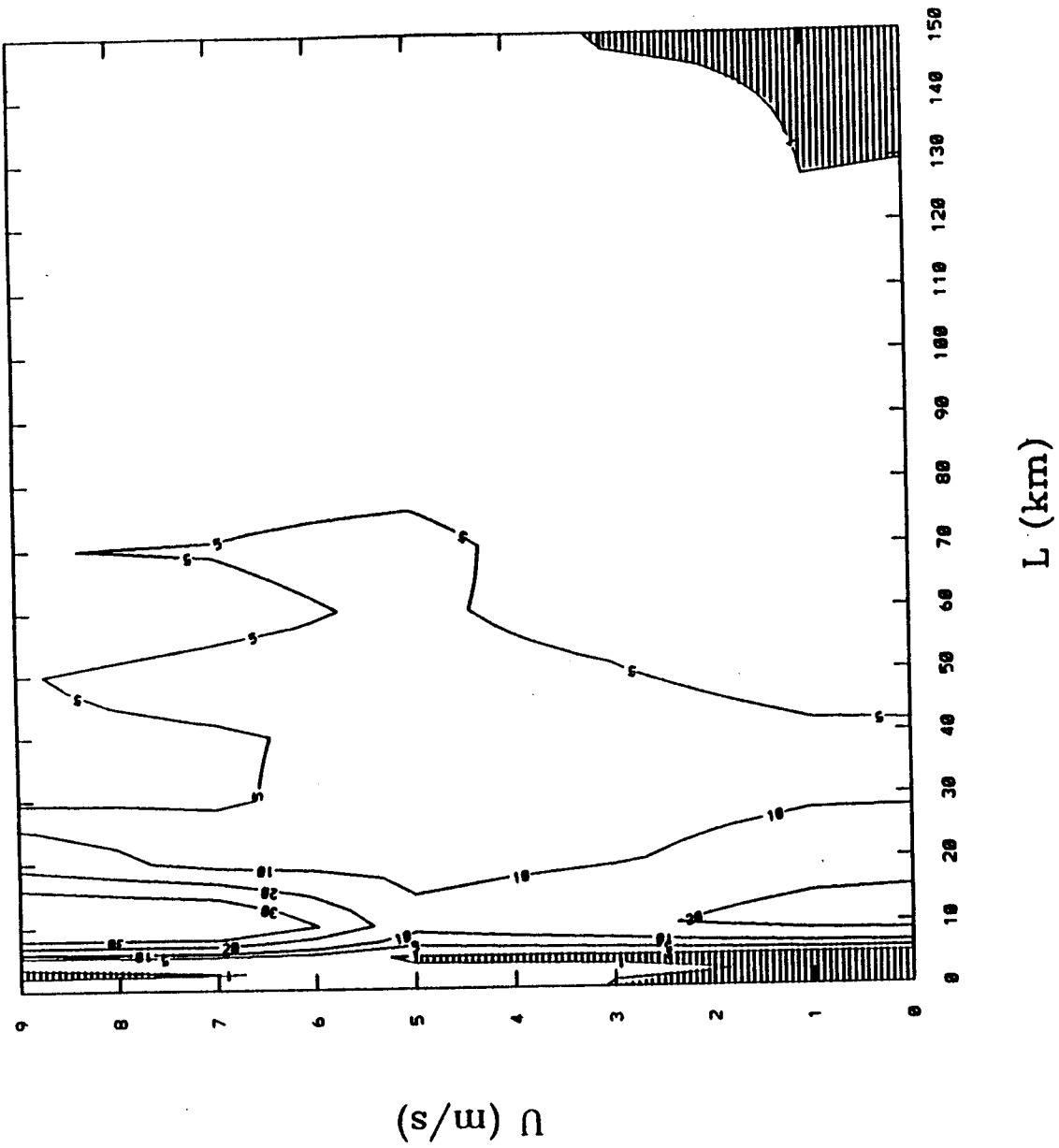
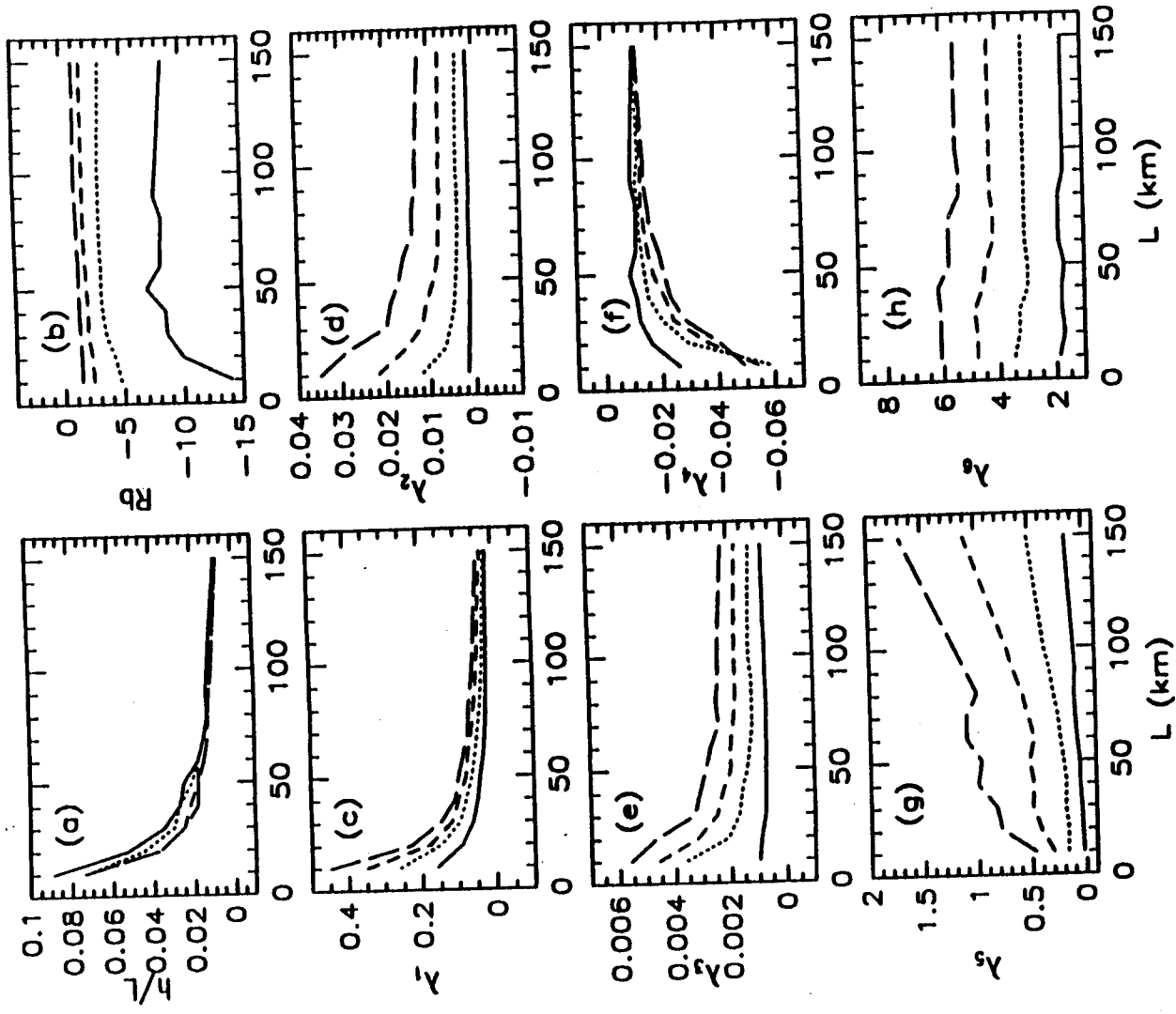


Fig. 4



7. List of Tables

Table 1: The parameters P_i used in Equation (2) for mesoscale [denoted by superscript " ' "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Table 2: The coefficients a_i used in Equation (2) for mesoscale [denoted by superscript " ' "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Table 3: The coefficients b_i used in Equation (2) for mesoscale [denoted by superscript " ' "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Table 4: The coefficients c_i used in Equations (2) and (3) for mesoscale [denoted by superscript " ' "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Table 5: The coefficients a_i , b_i , and c_i used in Eqs. (4) and (5) for total sensible heat flux scaled by its surface value.

Table 6: The coefficients a_i , b_i , and c_i used in Eqs. (4) and (5) for total moisture flux scaled by its surface value.

Table 7: The values of the coefficients a_i , b_i , c_i , and d_i used in Eqs. (7) and (8) for total momentum flux scaled by its surface value.

Table 8: The ratio of mesoscale to turbulent fluxes of heat, moisture, and momentum as a function of dimensionless height, z/z_i , for the FIFE study area. The parameter $L_x = 10$ km and other parameters are given in the text.

Table 9: Same as Table 8 except $L_x = 20$ km.

Table 1: The parameters P_i used in Equation (2) for mesoscale [denoted by superscript " ' "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (turbulent) surface values.

Parameters		P_1	P_2	P_3
dimensionless mesoscale flux	$< w'\theta' >$	λ_4	λ_6	z_i/L_x
	$< w'q' >$	λ_2	λ_4	λ_6
	$< w'u' >$	λ_1	z_i/L_x	R_b
dimensionless turbulent flux	$< w''\theta'' >$	λ_4	z_i/L_x	R_b
	$< w''q'' >$	λ_1	λ_3	z_i/L_x
	$< w''u'' >$	λ_4	z_i/L_x	R_b

Table 2: The coefficients α_i used in Equation (2) for mesoscale [denoted by superscript " / "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Coefficients		α_1	α_2	α_3	α_4	α_5	α_6
dimensionless mesoscale flux	$\langle w'g' \rangle$	2.57	249.20	2735.80	-0.202	0.209×10^{-1}	88.65
	$\langle w'q' \rangle$	7.35	597.70	-8507.40	322.70	3620.00	-2.20
	$\langle w'e' \rangle$	0.305	-80.64	137.10	461.10	-4762.10	0.724
dimensionless turbulent flux	$\langle w''g'' \rangle$	-0.908×10^{-2}	-4.01	-77.50	-2.32	23.30	0.597×10^{-2}
	$\langle w''q'' \rangle$	-0.523	-5.01	8.77	430.70	-73761.80	20.20
	$\langle w''e'' \rangle$	0.272×10^{-1}	16.90	169.60	11.50	-119.90	-0.365×10^{-1}
							0.34

Table 3: The coefficients b_i used in Equation (2) for mesoscale [denoted by superscript " / "] and small-scale (or turbulent) [denoted by superscript " " "] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Coefficients		b_1	b_2	b_3	b_4	b_5	b_6	b_7
dimensionless mesoscale flux	$\langle w'\theta' \rangle$	-3.66	-245.30	-2854.60	0.886	0.971×10^{-1}	-100.50	859.40
	$\langle w'q' \rangle$	-6.52	-434.20	6674.10	-221.20	-2620.00	2.49	-0.188
	$\langle w'u' \rangle$	-1.20	33.50	-118.60	-266.00	3641.50	-0.584	-0.545×10^{-1}
dimensionless turbulent flux	$\langle w'\theta'' \rangle$	-2.93	-272.30	-2731.50	-98.30	733.10	-0.364	-0.260×10^{-1}
	$\langle w''q'' \rangle$	1.08	48.80	-67.30	-1088.30	39871.10	-130.40	752.20
	$\langle w''u'' \rangle$	-4.58	-884.40	-8316.90	-386.40	3785.20	0.974	0.526×10^{-1}

Table 4: The coefficients c_i used in Equations (2) and (3) for mesoscale [denoted by superscript “'”] and small-scale (or turbulent) [denoted by superscript “''”] sensible heat, moisture, and momentum fluxes scaled by their corresponding total (or turbulent) surface values.

Coefficients		c_1	c_2	c_3	c_4
mesoscale flux	$< w'\theta' >$	0.00	1.30	-2.45	1.08
	$< w'q' >$	0.00	-0.101	4.50	-2.88
	$< w'u' >$	0.140×10^{-2}	-0.825×10^{-1}	0.562	-0.328
turbulent flux	$< w''\theta'' >$	1.00	-2.74	2.64	-0.905
	$< w''q'' >$	1.00	2.73	-5.20	1.87
	$< w''u'' >$	1.00	-3.84	5.12	-2.28

Table 5: The coefficients a_i , b_i , and c_i used in Equations (4) and (5) for total sensible heat flux scaled by its surface value.

Coefficients	a_1	a_2	a_3	a_4	a_5	a_6
$< w'\theta' > + < w''\theta'' >$	-0.546×10^{-1}	-10.20	92.10	3.98	0.410×10^{-2}	-0.367×10^{-2}
Coefficients	b_1	b_2	b_3	b_4	b_5	b_6
$< w'\theta' > + < w''\theta'' >$	3.22	289.00	-2593.90	-88.50	-0.387	0.211
Coefficients	c_1	c_2	c_3			
$< w'\theta' > + < w''\theta'' >$	1.0	-1.5	0.37			

Table 6: The coefficients a_i , b_i , and c_i used in Equations (5) and (6) for total moisture flux scaled by its surface value.

Coefficients	a_1	a_2	a_3	a_4	a_5	a_6	a_7	a_8
$< w'q' > + < w''q'' >$	0.939×10^{-1}	2.81	-6.51	8.26	134.70	-0.606×10^{-2}	0.586×10^{-4}	-9.72
Coefficients	b_1	b_2	b_3	b_4	b_5	b_6	b_7	b_8
$< w'q' > + < w''q'' >$	2.01	23.60	-29.10	65.60	400.70	-0.485	0.308×10^{-1}	-67.90
Coefficients	c_1	c_2	c_3					
$< w'q' > + < w''q'' >$	1.00	2.91	-1.80					

Table 7: The values of the coefficients a_i , b_i , c_i , and d_i used in Equations (7) and (8) for total momentum flux scaled by its surface value.

Coefficients	a_1	a_2	a_3	a_4
$< w'u' > + < w''u'' >$	-0.240	-8.19	0.264×10^{-1}	-2.49
Coefficients	b_1	b_2	b_3	b_4
$< w'u' > + < w''u'' >$	-1.61	-91.40	0.22	-34.80
Coefficients	c_1	c_2	c_3	c_4
$< w'u' > + < w''u'' >$	2.29	117.90	-0.334	49.90
Coefficients	d_1	d_2	d_3	d_4
$< w'u' > + < w''u'' >$	1.00	-3.97	6.00	-2.87

Table 8: The ratio of mesoscale to turbulent fluxes of heat, moisture, and momentum as a function of dimensionless height, z/z_i , for the FIFE study area. The parameter $L_x = 10$ km and other parameters are given in the text.

z/z_i	$\langle w'\theta' \rangle / \langle w''\theta'' \rangle$	$\langle w'u' \rangle / \langle w''u'' \rangle$	$\langle w'q' \rangle / \langle w''q'' \rangle$
0.1	0.21	0.015	0.0085
0.2	0.50	-0.040	0.031
0.3	0.89	-0.16	0.063
0.4	1.37	-0.37	0.11
0.5	1.93	-0.68	0.16
0.6	2.48	-0.91	0.23
0.7	2.71	-0.70	0.32
0.8	1.69	-0.19	0.45
0.9	-5.22	0.50	0.65
1.0	35.5	∞	1.02

Table 9: Same as Table 8 except $L_z = 20$ km.

z/z_i	$\langle w'\theta' \rangle / \langle w''\theta'' \rangle$	$\langle w'u' \rangle / \langle w''u'' \rangle$	$\langle w'q' \rangle / \langle w''q'' \rangle$
0.1	0.33	-0.0056	0.0085
0.2	0.77	0.024	0.037
0.3	1.34	0.15	0.087
0.4	2.06	0.49	0.16
0.5	2.92	1.33	0.27
0.6	3.81	2.79	0.42
0.7	4.27	4.07	0.64
0.8	2.76	4.77	0.97
0.9	-8.93	6.61	1.51
1.0	64.5	∞	2.52

The Influence of Topography on Meteorological Variables and Surface-Atmosphere Interactions

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Abstract: Topography perturbs practically every process and variable relevant to land-atmosphere interactions, including radiation, air temperature and saturation deficit, wind and turbulence, cloudiness and precipitation, and properties of the soil and vegetation. We review progress in understanding some of these influences, concentrating on the wind field and the exchanges of scalars (especially sensible and latent heat) over low to moderate hills, both without and with flow separation. The treatment of the wind field is based on a physical division of the flow into inner, outer and wake regions. This framework is used to discuss modelling approaches, measurements of both the mean flow and the turbulence, special processes in the wake region, and the effects of stratification.

Scalar transfer and the surface energy balance are discussed first in terms of a linear theory which accommodates radiative, aerodynamic, and elevation perturbations on the surface energy balance over hills. The linear theory provides general scales for determining the relative magnitudes of these perturbations, but its quantitative predictions are limited to low slopes without separation. Next, wind tunnel experiments are used to study the aerodynamic perturbations on scalar transfer caused by hills in situations outside the scope of the linear theory, such as flows with separation. Finally, the wind tunnel results are extrapolated to the atmosphere, using the framework of a Convective Boundary Layer model. It is found that, for relatively gentle topography, the regional surface energy balance (spatially averaged over several hills) is insensitive to the presence of the topography.

1. INTRODUCTION

Common observation shows that topography has a profound influence on the climate near the ground, and thereby on exchanges of energy, water and momentum between the land surface and the atmosphere. Hill crests are windier than valleys; hillslopes facing the equator are warmer than slopes facing the poles; topography strongly modulates cloudiness and precipitation; it is also significantly correlated with soil and vegetation properties. These influences occur through a wide range of processes, including dynamic (flow) effects of several kinds, thermodynamics, radiative effects, and soil and ecological effects. It is appropriate to begin by listing some of the major processes involved.

(1) *Flow of a nonstratified atmosphere around topography:* The flow around the topography itself has been the subject of intensive study. The picture depends significantly on whether the ambient atmospheric conditions are nonstratified (thermally neutral or unstable), or stably stratified; that is, whether or not the atmosphere exerts a restoring buoyancy force on a vertically displaced air parcel. In the case of nonstratified air flow over topographic obstacles, the flow sets up a pressure field which typically causes near-surface wind speeds to be greatest near summits and smallest in valleys. These changes in the mean wind field also affect the turbulence in several ways, particularly near the surface where stress gradients further influence the mean wind. In the lee of sufficiently steep topography, mean flow separation and flow reversal may occur. Whether or not there is mean flow separation, the wake region in the lee of a hill in a nonstratified ambient atmosphere is one of reduced wind speeds and high turbulence, particularly in the strong elevated shear layer which forms downwind of the crest of the hill.

(2) *Flow of a stratified atmosphere around topography:* The above dynamical picture is altered qualitatively if the ambient atmosphere is stably stratified, leading to several important differences from the neutral case. First, the flow over the hill can excite resonant, wave-like responses in the surrounding atmosphere, because the stable ambient atmosphere exerts a restoring buoyancy force on a vertically displaced air parcel (Smith, 1979; Durran, 1990). These responses typically induce an asymmetry in the flow near the hill, such that streamlines rise and decelerate (relative to the neutral case) upwind of the hill, and descend and accelerate on the lee side, sometimes leading into a train of "mountain waves" or "lee waves" (names often given to gravity waves excited by topography). Second, and in consequence, the point of maximum near-

surface wind speed moves down the lee side of the hill with increasing ambient stability. This can result in very low wind speeds on the upwind face and very high "downslope winds" on the lee side (Durran, 1990). Third, the gravity waves extract energy and momentum from the mean flow. Momentum extraction by this mechanism accounts for "wave drag", a significant loss term in the global momentum balance. The energy extracted into gravity waves can appear elsewhere in the form of turbulence (Einaudi and Finnigan, 1981; Finnigan and Einaudi, 1981), ultimately being dissipated as heat. Fourth, when conditions are very stable, the air may not be able to flow over a hill for energetic reasons, leading either to deviation of the flow around the hill, or (if this is impossible) to complete blockage of the flow (Sheppard, 1956; Smith, 1990).

The dynamical differences between the nonstratified and stratified situations lead to a practical meteorological distinction between low hills, large hills and mountains. Low hills have elevation changes which are only a small fraction of the depth of the daytime atmospheric boundary layer (ABL). The daytime ABL is normally nonstratified, being well mixed by convection, and is typically around 1 km deep, so low hills (in this sense) have elevation variations up to a few hundred metres. Therefore the daytime flow around such hills is usually nonstratified, or perturbed only slightly by stratification, though stratification and its consequences are significant at night. Large hills occupy a large fraction of the daytime ABL (typical elevation changes of around 1 km) and mountains protrude completely through daytime ABL. Topography of this magnitude induces significant perturbations in the stably stratified overlying tropospheric flow, so wind fields around large hills and mountains are almost always influenced heavily by the consequences of stratification.

(3) *Thermally forced topographic flows:* Of a different character to the above two flow categories are local, thermally driven circulations (Whiteman, 1990; Egger, 1990). These include katabatic and anabatic winds (downslope and upslope, respectively), and "valley winds" which blow up or down the longitudinal axis of a valley. The energy driving these flows is obtained not from the flow of the ambient atmosphere but from the interactions of local buoyancy fluxes with slopes. These flows are usually much weaker than those driven by the ambient atmospheric motion, except in extreme cases such as the katabatic wind off the Antarctic plateau. Thermally forced winds usually interact in subtle ways with the ambient atmospheric motion, even when this is weak.

- (4) *Thermodynamic processes*: The above three groups of processes all concern the wind field, but other classes of process are equally important in topographic modulation of the atmosphere. Of the thermodynamic effects, the simplest is adiabatic cooling and heating: in the absence of condensation and dynamically induced heat fluxes, air parcels cool at the dry adiabatic lapse rate (0.01 K m^{-1}) on ascent over a hill, and warm again in descent. The cooling is often sufficient to induce saturation and cloud formation, possibly leading to precipitation. Thus, cloudiness and precipitation are strongly correlated with topography, often more so than with any other controlling variable (Hutchinson, 1994). If the atmosphere is conditionally stable, the latent heat released by initial topographically induced condensation can lead to cumulus and thunderstorm development.
- (5) *Radiative processes*: The effects of slope on solar radiation are well known; for example, Paltridge and Platt (1976) review the relevant formulae. For small north-south oriented slopes at noon, the fractional change in direct solar irradiance with slope is approximately the product of the cotangent of the solar elevation and the sine of the slope, but the fractional change in net irradiance is somewhat less, because of the diffuse solar and longwave radiation components (Raupach *et al.*, 1992).
- (6) *Surface properties*: For land-atmosphere interactions, an additional group of important processes includes those governing the soil, hydrological and ecological properties of the surface. Soils are typically shallower, coarser-textured and less nutrient-rich on hill crests than in valleys. The soil water content is often strongly (though not usually simply) correlated with topography, first through the topographic influence on precipitation, and second because of lateral downslope movements of water after precipitation, both above and within the soil. These factors act together with modulations in physical microclimate (wind, temperature, radiation) to create ecological and vegetation changes with topography that can be as severe as those induced by latitudinal changes over continental distances. For instance, near the site of this meeting at Tucson, Arizona, the ecological variations over a 3000-metre range in altitude are similar to the variations along a 2000-km north-south transect from Mexico to Canada.
- (7) *Land-atmosphere exchanges*: All the above processes influence the land-atmosphere fluxes of energy, water, carbon and momentum. For example, let us consider the case of evaporation (a function of radiation, saturation deficit, wind speed and surface resistance) over terrain covered by long, parallel, sinusoidal ridges. For simplicity, we imagine that the ridgelines

are oriented normal to the direct solar beam throughout the day, with the sun passing directly overhead; that the ridges are low enough to be immersed in the ABL, so that the daytime ambient conditions are nonstratified; and that the wind is perpendicular to the ridges. For this situation, the topographically induced perturbations in some of the major meteorological and surface variables are sketched in Figure 1. Both the solar and net irradiances are greatest on the sunny slopes, and are therefore in quadrature with the elevation. The near-surface air temperature varies largely through adiabatic cooling and heating, and so is in antiphase with the elevation. Saturation deficit, responding mainly to air temperature, is also in antiphase. Since the flow is nonstratified, the perturbations in near-surface wind speed and surface stress are both nearly in phase with the elevation (maxima occur slightly upwind of crests). Assuming the surface to be vegetated, the surface resistance is likely to vary in phase with the elevation, because soil water deficits and plant water stresses are likely to be lowest in the valleys and highest on the crests. The effect of all of these differently-phased perturbations on the evaporation is clearly complex, and depends on which perturbation in forcing variables (radiation, humidity deficit, wind, surface resistance) dominates the evaporation perturbation. In a later section, calculations with a linearised model are used to unravel some of this complexity.

In the space of a single review, it is impossible to treat all of the above processes adequately. We therefore restrict attention to topics which are relevant for an assessment of topographic modulations to land-atmosphere exchanges; and even within this guideline, some important questions can only be treated briefly. Moreover, the selection of topics is inevitably biased towards work with which we are most familiar. The review is divided into three main sections, Section 1 being the above introductory survey. Section 2 deals with the changes to the velocity field (mean wind and turbulence) over topography, briefly covering a general theoretical framework, models, measurements, the special problems of the wake region, and stratified flows. Section 3 considers topographic effects on heat and mass transfer, with emphasis on the surface energy balance (SEB) in both local and spatially averaged forms. Our focus is mainly, though not exclusively, on the case of low hills as defined above.

2. WIND AND TURBULENCE OVER HILLS

2.1 Theoretical Framework

The study of turbulent wind flow over hills has proceeded rapidly in the last two decades, spurred by a combination of findings from field experiments, wind tunnel experiments, basic theory and numerical modelling. Recent reviews have been offered by Taylor *et al.* (1987), Finnigan (1988), Carruthers and Hunt (1990) and Kaimal and Finnigan (1994). To summarise the picture that has emerged, it is helpful to begin by dividing the flow into several regions with different dynamical properties. Figure 2 shows the main regions for nonstratified flow over a low hill: the flow on the upwind side and over the crest is divided into an *outer region* and an *inner region* (defined below), while on the downwind side of the hill there is a turbulent *wake region*, which may or may not involve flow separation. Figure 2 also introduces the main length scales, the hill height H and length L . The length L is defined for an isolated hill as half the distance between the upwind and downwind points where the elevation is $H/2$, and for repeated ridges as $(\text{terrain wavelength})/2$. Discussion of the wake region is mostly deferred to Section 2.4.

The division of the flow into inner and outer regions was introduced in a seminal paper by Jackson and Hunt (1975). The idea is that the velocity field is affected by the presence of the hill through two mechanisms: the pressure field around the hill, and changes in the turbulence, particularly the Reynolds stresses. This can be seen from the momentum balance for the flow of a neutral ambient atmosphere over a laterally uniform ridge normal to the mean wind. Retaining only the dominant terms, this is

$$U \frac{\partial U}{\partial x} + W \frac{\partial U}{\partial z} = - \frac{\partial P}{\partial x} + \frac{\partial \tau}{\partial z} \quad (1)$$

where x and z are Cartesian coordinates in the streamwise and vertical directions, respectively; U and W are streamwise and vertical mean velocity components; P is mean kinematic pressure; and $\tau = -\overline{uw}$ is the Reynolds shear stress, with u and w the fluctuating velocity components.

When the flow goes over the hill, a perturbation is set up in the pressure field $P(x,z)$, extending over horizontal and vertical distances which are both of the same order of magnitude as the hill

length L , as sketched in Figure 3a. The $\partial P/\partial x$ term in Equation (1) therefore perturbs the velocity field throughout over a region with depth of order L . On the other hand, the $\partial\tau/\partial z$ term perturbs the velocity field only in the region through which perturbations in stress (τ) can diffuse. Figure 3b sketches a typical pattern for the τ perturbations: they originate at the surface (in the form of alterations in the surface shear stress following the mean wind field) and diffuse upwards with a transport velocity determined by the turbulence itself. A suitable scale for the order of magnitude of this transport velocity is u_* , the friction velocity in the undisturbed, background flow far upwind of the hill. Then, if U_b is the background mean flow velocity, stress perturbations reach a height of order $L u_* / U_b$ in the time it takes for the flow to travel over the hill. This is a layer much shallower than L , because u_* / U_b is small (typically 0.05 to 0.1). The *outer region* is the part of the flow where only the pressure gradient is important in modifying the velocity field, while the *inner region* is the shallower layer where the velocity field is modified by both pressure and stress gradients.

The depth l of the inner region is usually regarded as a global parameter for flow over a hill, although the inner region as defined above is not of uniform depth (Figure 3b). Jackson and Hunt (1975) estimated l as the height at which the perturbation stress divergence and advection terms in Equation (1) are of the same order of magnitude. For neutral flow with a logarithmic upwind velocity profile, this leads to the implicit equation

$$l/L = 2\kappa^2 \ln^{-1}(l/z_0) \quad (2)$$

where z_0 is the roughness length and κ the von Karman constant. The resulting values of l are of the same order as the above rough estimate $l = L u_* / U_b$, if U_b is evaluated at height l . Typically, l is in the range 0.05 L to 0.1 L , increasing with z_0 .

The division of the flow into inner and outer regions has two key consequences. The first is that the outer region is one where the response of the mean flow to the hill is inviscid; that is, independent of changes in the turbulence because the effect on the mean flow of changes in turbulence is confined to the inner region. In the outer region, the equations of motion and continuity can be cast in perturbation form, written in the inviscid approximation (by neglecting

terms involving turbulent stresses), and linearised (by neglecting perturbation product terms). The flow is then governed by a single equation for W , which in the two-dimensional case is ,

$$\nabla^2 W - \frac{U_b''(z)}{U_b(z)} W = 0 \quad , \quad (3)$$

where $U_b(z)$ is the background or upwind velocity profile; $\nabla^2 = \partial^2/\partial x^2 + \partial^2/\partial z^2$; and primes denote differentiation in z . With boundary conditions specified by the hill geometry, this equation can be solved for $W(x,z)$ in the outer region, and thence (via the equations of continuity and motion) for $U(x,z)$ and $P(x,z)$. Hunt *et al.* (1988a, henceforth HLR) introduced a further division of the outer region into an *upper layer* in which the upwind shear dU_b/dz is negligible, so that the flow is irrotational as well as inviscid and Equation (3) simplifies to the Laplace equation $\nabla^2 W = 0$; and a *middle layer* in which dU_b/dz cannot be neglected, giving a rotational but inviscid flow satisfying the complete Equation (3). The dividing height, or middle-layer depth h_m , satisfies the implicit equation $h_m/L = \ln^{-1/2}(h_m/z_0)$; this gives typical values $h_m \approx L/3$ (compared with $l \approx L/10$, from Equation (2)).

The second key consequence is that it is reasonable to take the pressure field $P(x,z)$ as constant with height ($\partial P/\partial z = 0$) through the inner region, because $P(x,z)$ has a vertical length scale of variation (L) which is much greater than the inner-region depth l . Therefore, the flow in the inner region obeys an equation of motion like Equation (1), but with the *imposed* pressure field $P(x,z) = P_0(x)$, where $P_0(x)$ is the hill-surface pressure field from the outer-region solution following from Equation (3). The problems of finding the pressure field and calculating changes in turbulence over the hill are thereby partly decoupled, offering major benefits in computational simplicity in addition to conceptual insights.

We turn briefly from the mean flow to the turbulence, which responds to the topography in qualitatively different ways in the outer and inner regions. The upwind turbulence can be characterised by a height-dependent time scale, $T_e = \epsilon/\epsilon$, for the "lifetime" of a typical eddy at height z (where ϵ is the turbulent kinetic energy and ϵ its dissipation rate at height z). In a

neutral surface layer, T_e is of order z/u_* . In the outer region, T_e exceeds the advection time over the hill (L/U_b), while in the inner region, $T_e < L/U_b$. In fact, the height where $T_e = L/U_b$ provides an alternative definition of the inner-region depth l , equivalent to Equation (2) to within an $O(1)$ constant (Kaimal and Finnigan, 1994). It follows that only the turbulence very close to the ground (at heights $z \ll l$) can reach local equilibrium with the continually changing mean velocity field as the flow passes over the hill. The flow in this very shallow layer is related to the familiar neutral surface layer over level terrain in which the turbulence is in local equilibrium with a logarithmic mean velocity profile (Townsend, 1976; Raupach *et al.*, 1991), but the nature of the equilibrium on the hill is modified by the flow acceleration and curvature (Kaimal and Finnigan, 1994). In the outer region, the turbulence can never equilibrate to local conditions, but rather responds "rapidly" (that is, on time scales $\lesssim T_e$) to mean flow changes including acceleration, shear and curvature. This response can be determined from linear "rapid-distortion theory" (Hunt and Carruthers 1990), in which the effects of mean flow changes on the turbulence are calculated from linear equations, ignoring the nonlinear turbulence-turbulence interactions because they require time scales $\gtrsim T_e$ to take effect.

2.2 Models

Several kinds of model are available to predict wind flow over hills (and also scalar transfer, discussed later). Multilayer models use the above framework explicitly, thereby gaining great computational efficiency. Other kinds of model compute rather than impose the layer structure, gaining generality at the cost of computational demand. The following summary covers five of the principal modelling approaches.

- (1) *Linear multilayer models* are based on the distinction between the outer and inner regions, and the associated decoupling of pressure and turbulence influences. In summary, the approach is: (a) flow equations are written for two (or more) regions, retaining different terms in each region as above; (b) these equations are then linearised, by expressing each mean flow variable as the sum of an undisturbed, background component and a perturbation component due to the hill; (c) linear equations are formed for the perturbation components, neglecting perturbation product terms relative to linear terms, which is a good approximation if the perturbations are small enough; (d) the linearised equations are then solved in the various layers, with

asymptotically matched boundary conditions. Equation (3) is an example of a linearised equation for the outer region. Linearisation simplifies the computational task enormously, but carries two main penalties: first, linear theories apply exactly only to hills of infinitesimal slope H/L , although in practice they work reasonably well up to H/L around 0.3 before the neglect of the nonlinear terms becomes crucial. Second, linear models cannot predict separation or near-separation in the wake region, because these processes are highly nonlinear, involving mean flow perturbations comparable with the background flow, and also strong turbulence-turbulence interactions. Therefore, linear models generally work much better on the upslope and crest of a hill than on the downslope. Semi-analytic examples of linear models include Jackson and Hunt (1975) and HLR for nonstratified flow; Hunt *et al.* (1988b) for stably stratified flow; and Hunt *et al.* (1988c) and Raupach *et al.* (1992) for scalar transfer. Other authors have concentrated on the use of improved turbulence closures, particularly Sykes (1980), Newley (1985), Zeman and Jensen (1987).

(2) *Mixed Spectral Finite Difference (MSFD) models* use linearised flow equations, solved by spectral methods in the horizontal and finite-difference methods in the vertical. However, unlike the previous class of models, they do not use analytical results specific to various flow regions (although use is made of the general form of the analytical solutions to simplify upper boundary conditions and stretch the vertical grid appropriately). The first MSFD model was developed by Beljaars *et al.* (1987) for the velocity and shear stress fields; extension to scalar fields was made by Padro and Walmsley (1990). The turbulence closure was extended to higher orders, including normal stresses, by Ayotte *et al.* (1994). A neutral planetary boundary layer version has been developed by Ayotte and Taylor (1995). Ayotte (work in progress) has also developed an adjoint method for data assimilation into the zero-order solution about which perturbations are constructed.

(3) *Nonlinear models*, based on the MSFD approach, have recently been developed. These relax the heavy restriction of linearity, thus accommodating large perturbations of the order of the mean flow (though there are still restrictions on the slopes for which these models are applicable). High computational efficiency is achieved by using nonlinear terms as sources in an iterative solution method. Xu and Taylor (1992) lead this effort through NLSFD, a nonlinear version of MSFD, extended to incorporate higher-order turbulence closures by Xu *et al.* (1994).

(4) *Finite-difference models* are fully nonlinear, solving the Reynolds equations on a three-dimensional grid, but need much more computer power than multilayer models. Finite-difference models must be nonhydrostatic to predict the effects discussed in this paper. Early finite difference models of flow over hills were constructed by Taylor (1977a,b). Also in this class are modern comprehensive mesoscale models such as RAMS (Pielke *et al.*, 1992) which account in principle for practically all the topographic processes mentioned in Section 1. These are very powerful, but their application to the problems discussed in this paper is still at an early stage.

(5) *Large-eddy simulation* (LES) uses intensive computation to solve the equations of motion in both space and time, on a grid fine enough to resolve the complete eddy structure except for the smallest scales. The first atmospheric LES studies were on the convective ABL (e.g. Moeng, 1984), but recent LES applications have spanned many important ABL problems including the stable ABL, patchy heat flux distributions, flows in canopies and flows around obstacles. In our context, Walko *et al.* (1992) and Dörnbrack and Schumann (1993) have used LES recently to study convective flow and heat transfer around sinusoidal ridges. Although computer-intensive and not for routine use, LES is a very powerful research tool which is likely to be applied increasingly to topographic flow and transfer problems in the next few years.

2.3 Observations

Over the last 15 years, a sequence of major field experiments has established the main phenomenological properties of near-neutral, shear-driven flow over low hills. Some of the main experiments are listed in Table 1; for a more extensive list, see Taylor *et al.* (1987). These atmospheric experiments have been backed up by extensive laboratory experimental work in wind tunnels (for example, Britter *et al.*, 1981; Teunissen and Shokr, 1985; Gong and Ibbetson, 1989; Finnigan *et al.*, 1990; Finnigan and Brunet, 1994). Here we summarise the basic picture that has emerged for both the mean velocity field and the turbulence.

When discussing measurements, the issue of coordinate frameworks becomes critical. We have so far assumed conventional rectangular Cartesian coordinates, but a more logical alternative for hill flows is *physical streamline coordinates* (Finnigan, 1983, 1990; Kaimal and Finnigan, 1994) in which the x and z axes are locally parallel and normal to mean streamlines. Although the coordinate framework is locally rectangular, the coordinate directions change from place to place. Its advantages are, first, that the ascent and descent of the flow over the hill are

incorporated directly into the coordinates, with the mean vertical wind (W) being replaced as a flow variable by streamline radius of curvature (R). The dynamical effects of streamline curvature on both the mean flow and the turbulence are thereby made more explicit. Second, from a measurement viewpoint, turbulent fluxes like $\overline{uw}$ represent purely cross-stream fluxes in streamline coordinates, instead of awkward mixtures of along-stream and cross-stream components. The penalty for using streamline coordinates is that the flow equations acquire extra terms; however, many of these terms are small and can often be ignored if the hill is not too steep. The equation of mean motion in the streamwise direction, for example, still takes the form of Equation (1) (without the W advection term) in streamline coordinates, if only leading-order terms are retained and $H/L \ll 1$. In the following discussion of measurements, a streamline coordinate framework is assumed unless otherwise specified.

Figure 4 summarises some of the main features of the mean velocity field over a two-dimensional (2D) ridge normal to the mean wind, and over an axisymmetric three-dimensional (3D) hill. At the upwind foot of the 2D ridge (but not the 3D hill), a slight deceleration is observed in the near-surface (inner-region) flow. This is in response to the decelerating (positive) surface pressure gradient $\partial P/\partial x$ (see Figure 3a). The flow then accelerates to the hill top, driven by a strong accelerating (negative) pressure gradient. In the wake region on the downwind side of the hill, strong deceleration is observed. Figure 5 demonstrates these features by showing observed profiles of mean velocity $U(z)$ for flow over a typical hill with H/L around 0.3, at x locations far upwind of the hill, at the hill crest, and in the wake region. The curves are abstracted from the field and wind tunnel observations mentioned above. The height axis is logarithmic, with z normalised by the inner-region depth l .

The effect of topography on the mean wind field can be described conveniently by the change in mean velocity at a particular height, $\Delta U(x, z) = U(x, z) - U_0(z)$. As indicated in

Figure 5, ΔU has a bow-shaped profile with height z at the hill crest. Of practical importance are the height z_{max} of the maximum in ΔU at the crest, and the magnitude ΔU_{max} of this

maximum. The linear theory of HLR predicts that $z_{max} \approx l/3$, where l is the inner-region depth,

and that

$$\Delta U_{max} \approx \frac{H}{L} \left[\frac{U_b^2(h_m)}{U_b(l)} \right] \left[1 + \frac{1.8}{\ln(l/z_0)} \right] \quad (4)$$

where h_m is the middle layer depth defined above. These predictions are in reasonable accord with most of the observations in Table 1. A closely related indicator of mean velocity changes is the fractional speedup $\Delta s(x, z) = \Delta U(x, z)/U_b(z)$, or ΔU normalised by the background mean velocity at the same height above ground. The speedup is predicted by HLR to be of the form

$$\Delta s(x, z) = \frac{\Delta U(x, z)}{U_b(z)} = \frac{H}{L} \left[\frac{U_b^2(h_m)}{U_b(l)U_b(z)} \right] \eta(x, z, z_0) \quad (5)$$

where $\eta(x, z, z_0)$ is a function of order 1, dependent on the precise hill shape and the upwind roughness length z_0 (details are given in HLR). As a test of the HLR prediction, Figure 6 shows a transect with x of measured speedups at height 2 m on Cooper's Ridge, New South Wales (Coppin *et al.*, 1994). The HLR prediction is shown both for the actual hill shape and also for the "Witch of Agnesi" shape, a commonly used analytical shape for a 2D ridge, given by

$$Z(x) = \frac{H}{1 + (x/L)^2} \quad (6)$$

where $Z(x)$ is the local elevation. This hill shape has convenient mathematical properties, and is the shape sketched in Figures 2 and 4. The predictions in Figure 6 for the two shapes are fairly similar, indicating that the speedup is not grossly sensitive to details of the hill shape, and are in some agreement with the measurements. The HLR theory usually works reasonably well on the upslopes of hills in nonstratified flow, but fails in the wake region where linear theories tend to break down.

Turning to the turbulence, Figure 7 shows typical observed profiles in the standard deviations of the streamwise, lateral and vertical wind components, σ_u , σ_v , and σ_w , and the Reynolds shear stress $\overline{uw}$, at the crest of a low hill with H/L around 0.3, immersed in a deep turbulent boundary layer. Each quantity is normalised with its value in the background, upwind flow at the same height above local ground, and the height axis is logarithmic and normalised with l . Figure 7, like Figure 5, is a synthesis of findings from the field experiments in Table 1, together with wind tunnel data. The rather complex behaviour of the profiles shown in Figure 7 occurs as the turbulence strikes a balance between the competing demands of equilibrating with changes in the mean flow at $z \ll l$ (where the turbulence time scale $T_e \ll LU_b$, the travel time over the hill) and responding "rapidly" to shear, acceleration and curvature changes in the mean flow at $z \geq l$ (where $T_e \geq LU_b$). Around $z \approx l$, all of these effects tend to be of the same order and it is impossible to predict the form of the turbulence profiles without using a model with sophisticated turbulence closure, though Kaimal and Finnigan (1994) give a heuristic explanation of the behaviour of the stress field through its relationship to the mean vorticity field. One important consequence of the interplay between turbulence and the mean flow is that the surface drag coefficient (defined as $\tau(z=0)/U^2(z=z_{max})$) varies over the hill. On a isolated 2D ridge with $H/L = 1/3$, the drag coefficient at the hill crest is about 60% of its upwind value (Finnigan *et al.*, 1990).

If the hill is covered with tall roughness, where the canopy height is a significant fraction of the inner region depth l , then the picture we have just sketched is changed substantially in the inner region. The near-surface region of local equilibrium is now replaced by a plant canopy flow, modified from its usual, level-terrain form (e.g. Raupach, 1988) by flow acceleration and

curvature. Finnigan and Brunet (1994) made a wind tunnel study of a model canopy (height h_c) covering a 2D ridge with height $H = 3h_c$ and length $L = 8.4h_c$, finding a pronounced variation over the hill of the depth of the characteristic layer of strong shear and turbulence near the canopy top, which deepens around the lower third of the upwind slope, and becomes much shallower near the hill crest. Such modification of the flow and turbulence in the canopy by topography is likely to be important for scalar transfer processes, especially for those dominated by aerodynamics such as particle deposition.

Most of the features sketched above apply to isolated hills. There is not yet a comparable body of experimental data for flow over multiple hills; the few available studies include the field experiment of Bradley *et al.* (1986), the wind tunnel study of Gong and Ibbetson (1989) and the numerical model comparison of Xu *et al.* (1994). These indicate that after many similar hills have been traversed, the mean velocity perturbation converges to a smaller value than it would have over the same hill in isolation, whereas turbulence perturbations converge to larger values.

2.4 The Wake Region

For nonstratified flow, the wake region is one of low wind speeds and high relative turbulence intensities (σ_v/U). If the hill slope behind the crest is steep enough, then a region of recirculating, separated flow will develop. This has significant consequences for the transfer of momentum and scalars, because the separated region is largely decoupled aerodynamically from the flow above. Therefore, it is important to be able to predict when separation will occur and how large the separated region is. Unfortunately, there is no simple theoretical way of making these predictions, so we are forced to rely on rules of thumb generated from experience. From a review of many field and wind tunnel experiments, Finnigan (1988) suggested the following picture for nonstratified hill flows: the occurrence of separation in the wake region depends on hill shape (2D or 3D), steepness and roughness. Steepness always enhances separation; for hills of a given steepness, separation is more likely with increasing roughness; and separation is more likely over 2D ridges than over axisymmetric 3D hills, other things being equal. In quantitative terms, separation is likely to occur when the maximum hill slope on the lee side, θ , exceeds the values given in Table 2.

Whether a separation region is formed or not, a wake region of reduced mean velocity and enhanced turbulence extends for a considerable distance downwind of the hill. The properties of the flow in this region have much in common with flows around obstacles such as buildings and windbreaks, which were reviewed comprehensively by Taylor (1988). The mean velocity and turbulence structure in the near wake (say within a few L of the crest) depend strongly on the details of the hill shape. In the far wake, well downwind of the hill, there is some evidence that the flow has "self-preserving" properties (that is, there are height and velocity scales, with power-law dependences on x , which collapse the observed flow perturbations to universal curves). However, the available information "has a distinct signal-to-noise problem" (Taylor, 1988), and the nature and even existence of these self-preserving properties is not yet resolved.

2.5 Effects of Ambient Stratification

As indicated in the introduction, the effects of stratification are profound. For stably stratified flow, a measure of the stratification is provided by the dimensionless *Froude numbers* $F_L = U/NL$ and $F_H = U/NH$. Here N is the Brunt-Vaisala frequency in the stable ambient atmosphere, defined by $N = [(g/\theta)(\partial\theta/\partial z)]^{1/2}$, with g the gravitational acceleration and θ the mean potential temperature; N is the angular frequency of vertical oscillation of an air parcel under the restoring force of buoyancy. The Froude numbers F_L and F_H , based respectively on hill length and height, are infinite in a neutral ambient atmosphere ($N \rightarrow 0$), and decrease towards zero with increasing stability ($N \rightarrow \infty$). Since both U and N are height-dependent, values at L are used for F_L , and at H for F_H . Kaimal and Finnigan (1994) classified stability classes for stratified flow over hills as:

Weakly stable: $F_L > 1, F_H \gg 1, u/NL \lesssim 1$

Moderately stable: $F_L \lesssim 1, F_H > 1$

Strongly stable: $1 \gtrsim F_H > 0$

In comparison with the nonstratified flows discussed above, the available base of experimental information on stratified hill flows is small, so it is necessary to rely more heavily on theory and models. In weakly and moderately stable flows, it is still appropriate to divide the flow into an inviscid-response outer region and an turbulent-response inner region, with the pressure field $P_0(x)$ in the inner region set by the surface pressure from the outer-region solution. Moreover, it is still useful to use linear theory to describe the flows in the two regions, provided the steepness H/L is not too great (Hunt *et al.*, 1988b). For the outer region, the linearised governing equation in 2D flows is a generalisation of Equation (3):

$$\nabla^2 W + \left[\frac{N^2(z)}{U_b^2(z)} - \frac{U_b''(z)}{U_b(z)} \right] W = 0 \quad (7)$$

in which the first term in square brackets accounts for the effects of buoyancy and the second for shear. The consequences of this equation can be glimpsed by examining the case of constant, height-independent N and U_b , for which analytic solutions are available. Figure 8 shows the inner-region mean velocity change, $[U(x,z) - U_b(z)]/U_b(h_m)$, normalised with the upwind velocity at the middle-layer height h_m defined earlier, calculated using such a solution augmented with an assumption that the background $U_b(z)$ is constant above h_m and varies below h_m in accord with standard Monin-Obukhov similarity theory (Kaimal and Finnigan, 1994). In nonstratified flow, the speedup is nearly symmetric about the hill crest (recalling that the strong velocity reduction usually seen in the wake region is a nonlinear phenomenon, invisible to this kind of linear theory). With increasing stability (decreasing F_L), the maximum velocity amplifies and moves its location from just upwind of the crest to near the bottom of the lee side of the hill, implying strong downslope winds (Durrant, 1990). Lee wave oscillations also begin to appear.

Strongly stable flows are too heavily perturbed by stability for the linearised Equation (7) to be valid, even for small H/L . The pressure field is determined mainly hydrostatically, in contrast to neutral and weakly stratified situations where the pressure field is dominated by the dynamics through Equation (3) or (7). When $F_H < 1$, the flow may not have sufficient kinetic energy to pass over the hill and is then forced to flow around a 3D hill, or becomes blocked by a 2D hill. This occurs below a "dividing streamline" at height h_s , which can be estimated as $h_s = H(1 - F_H)$ in the simple case of constant U_b and N (Sheppard, 1956). A critique and

substantially revised physical view of this concept was presented by Smith (1990), but the simple formula compares surprisingly well with towing tank data (Snyder *et al.*, 1985).

3. SCALAR TRANSFER OVER HILLS

There is a significant body of work on scalar transfer around hills in the context of the behaviour of contaminant plumes from localised sources; see Hanna and Strimaitis (1990) for a review. Unfortunately, there is little available work on the surface exchange processes that concern us here. On the experimental side, Verma and Cermak (1974) measured the temperature field over a heated ridged surface in a wind tunnel; Dawkins and Davies (1981) reported measurements of evaporation (of aniline) over a low ridge in a wind tunnel; and field measurements of turbulent heat flux over a low ridge were made by Coppin *et al.* (1994) as part of an experiment mainly focussed on the wind field. This experimental work is indicative rather than comprehensive. On the theoretical side, an attempt at a systematic approach to the effect of hills on the surface energy balance (SEB) is the linear analytic model of Raupach *et al.* (1992, henceforth RWCH) which is based on the linear analysis of HLR for the wind field, and its extension to scalars by Hunt *et al.* (1988c). The linear approach restricts the quantitative predictions of this analysis to low hills without separation. However, the theory also provides a general assessment of the several processes affecting scalar transfer, and their scaling with external controlling parameters such as hill slope H/L , ambient irradiance, wind speed, saturation deficit and surface properties. To this extent, the linear analysis is useful outside its formal bounds of applicability.

This section therefore consists of three parts: a description of the processes and parameters acting on the SEB in hilly terrain, based on the linear RWCH analysis; a description of a wind tunnel experiment on scalar transfer over hills, in conditions outside the strict limits of the linear theory; and a scaling-up exercise, in which the results of the wind tunnel experiment are translated into the field with the aid of a CBL slab model.

3.1 *Linear Analysis of the Surface Energy Balance in Hilly Terrain*

The linear analyses of Hunt *et al.* (1988c) for scalar transfer, and RWCH for coupled sensible and latent heat transfer, divide the flow into inner and outer regions in the same way as

HLR for the velocity field. However, a key distinction between the scalar and velocity fields is evident by comparing the momentum conservation equation, (1), with the corresponding scalar equation. For an arbitrary conserved scalar (such as temperature, humidity or other atmospheric constituent) with mean concentration $S(x,z)$ and vertical flux $F(x,z)$, this is:

$$U \frac{\partial S}{\partial x} + W \frac{\partial S}{\partial z} = - \frac{\partial F}{\partial z} \quad (8)$$

Because there is no pressure term here, perturbations to the scalar field (arising from changed surface boundary conditions) can only be translated into the body of the flow by turbulent transfer, at an upward velocity of order u_* . Therefore, the main effects of hills on S and F are confined to the inner region, the outer region being affected only through the vertical displacement of mean streamlines. The inner region depth for scalars can still be estimated from Equation (2).

To solve Equation (8) in the inner region, Hunt *et al.* (1988c) first related F to S with the gradient-diffusion assumption $F = -K \partial S / \partial z$, with the eddy diffusivity K proportional to $\tau^{1/2} z$, where $\tau(x,z)$ is the local kinematic shear stress. Using the theory of HLR to determine the perturbations in U , W and K , they obtained and solved a linear equation for perturbations in S . The solution shows that S perturbations arise through three aerodynamic mechanisms: (a) convergence and divergence of mean streamlines; (b) vertical stress gradients $\partial \tau / \partial z$, which cause changes in K ; and (c) changes in boundary conditions. Through linearity, the solution for S perturbations consists of three additive terms, one for each mechanism. In practice, the S perturbations near the surface ($z \lesssim l$) are dominated by changes in the boundary conditions, with the other two mechanisms making fairly small contributions of opposite sign.

In the case of coupled heat and water vapour transfer there are two scalars, the potential temperature θ and specific humidity Q . Their surface boundary conditions are *coupled*, because changes in the θ field affect the surface flux of Q (via the saturation deficit), and *vice versa*. Before the problem can be analysed for each scalar individually, these boundary

conditions must be decoupled. Following McNaughton (1976), a general way to do this is to define two new scalars, each a linear combination of Θ and Q :

$$\begin{aligned} A &= \rho c_p \Theta + \rho \lambda Q \\ B &= \epsilon_0 \rho c_p \Theta - \rho \lambda Q \end{aligned} \quad (9)$$

where ρ is air density, c_p isobaric specific heat of air, λ latent heat of vaporisation of water, and $\epsilon_0 = (\lambda/c_p)(dQ_{sat}/dT)$, with $Q_{sat}(T)$ the saturation specific humidity (ϵ_0 is the ratio of latent to sensible heat increase with temperature in saturated air, about 2.2 at 20 C). The scalar A is the available energy density or enthalpy per unit volume, and B is a linearised form of the saturation deficit in energy density units. Both A and B obey "private", decoupled boundary conditions. Replacing Θ and Q with A and B , it is possible to analyse a range of advection-dispersion problems involving coupled heat and water vapour transfer, including convective boundary layers over heterogeneous surfaces (Raupach, 1991) and energy balances in vegetation canopies (Raupach, 1988; Dolman and Wallace, 1991). In the present case, RWCH determined A and B from the solution of Hunt *et al.* (1988c) for an arbitrary scalar, and thus found the spatial distribution of Θ and Q (and the corresponding heat and water vapour fluxes) over low hills.

The analysis leads to four general conclusions:

- (1) Perturbations in the temperature and humidity fields and fluxes arise from four physical causes: *radiative effects* due to slope, causing changes in available energy; *aerodynamic effects* (including changes in the surface stress, vertical stress gradients and streamline convergence and divergence); *elevation effects* induced by adiabatic cooling on ascent; and changes in *moisture availability* at the surface, described (for instance) by a bulk surface resistance. For simplicity, RWCH ignored the last effect, assuming the bulk surface resistance to be constant over the hill.
- (2) Apart from moisture availability, these processes are governed by three external dimensionless parameters: the hill slope H/L ; the upwind, background evaporative fraction α_1 , $= F_E/F_{A1}$ (with F_E the latent heat flux, F_A the available energy flux or net irradiance less heat flux

into thermal storage, and the subscript 1 denoting a background value); and a dimensionless elevation parameter P_{elev} , proportional to the adiabatic change in saturation deficit through height H .

(3) These processes are not always equally important. For example, let ϕ_E be the fractional change in the surface latent heat flux F_E on the hill, normalised with the background flux F_{E1} , so that $F_E = F_{E1}(1 + \phi_E)$. Contributions to ϕ_E from radiative effects scale with $(H/L)\alpha_1^{-1}$; contributions from aerodynamic effects scale with $(H/L)(\alpha_{eq}^{-1} - \alpha_1^{-1})$; and contributions from elevation effects scale with P_{elev} (RWCH, Equation (4.10)). Here, $\alpha_{eq} = \epsilon_0/(\epsilon_0 + 1)$ is the evaporative fraction for equilibrium evaporation, about 0.68 at 20 C (Slatyer and McIlroy, 1961; Thom, 1975; Raupach, 1994). These scaling factors imply that radiative and aerodynamic effects are both proportional to hill slope H/L , whereas elevation effects are proportional to hill height H . Also, aerodynamic effects make a positive contribution to the latent heat flux perturbation ϕ_E when α_1 exceeds α_{eq} , a negative contribution when $\alpha_1 < \alpha_{eq}$, and zero contribution when $\alpha_1 = \alpha_{eq}$ (where a "positive contribution" means a perturbation in phase with the topography, positive on crests and negative in valleys, while a "negative contribution" is in antiphase with the topography). This is consistent with the fact that, for thermodynamic reasons, equilibrium evaporation is the point at which the evaporative fraction is independent of both wind speed and surface roughness (Thom, 1975). Therefore, when $\alpha_1 = \alpha_{eq}$, aerodynamic effects induce no evaporation perturbations (though other processes still operate). Because evaporation in nature is often quite close to the equilibrium value, heat and water vapour transfer (and thence the SEB) are less sensitive to aerodynamic influences than other forms of land-air exchange such as momentum transfer and the dry deposition of pollutants, where no equivalent thermodynamic constraint exists.

(4) The linear RWCH theory predicts the spatially averaged flux perturbation $\langle \phi_E \rangle$ over a hilly region, finding that contributions to $\langle \phi_E \rangle$ from radiative and aerodynamic effects are zero, while elevation effects make a small contribution due to adiabatic changes in saturation deficit. Thus, although scalar fluxes (including the sensible and latent heat fluxes) undergo significant local perturbations because of the topography, the spatially averaged fluxes remain very close to background values. On one level, this result is a trivial consequence of linearisation (only the DC component can influence mean values in a linear theory) but the point is that, insofar as linearisation is valid, this conclusion is an immediate consequence.

As an example of the predictions of the RWCH theory, Figure 9 shows the perturbations in the surface latent heat flux F_E , sensible heat flux F_H and available energy flux $F_A = F_E + F_H$, over a "Witch of Agnesi" hill (Equation (6)) with $H = 100$ m and $L = 500$ m, and with the sun shining at elevation 45° on the upslope (see RWCH for other details). Three cases (A, B and C) are shown, with decreasing background available energy fluxes ($F_{A1} = 500, 250$ and 10 W m^{-2} , respectively) but all other conditions identical. Table 3 shows the resulting background SEB: the evaporative fraction α_1 is very close to the equilibrium value ($\alpha_{eq} = 0.686$) for case A, larger for case B and very large for case C, in which the available energy is almost zero and the latent heat flux is sustained by a downward sensible heat flux.

The available energy perturbations on the hill (Figure 9) are due entirely to slope effects on radiation, so they are positive, zero and negative on the upslope, crest and downslope, respectively. In case A, radiative effects are the main contributors to perturbations in F_E and F_H ; aerodynamic effects are negligible because $\alpha_1 \approx \alpha_{eq}$, and elevation effects are fairly small for a hill of height 100 m. Accordingly, perturbations in both F_E and F_H follow the available energy perturbation closely. Case C represents an opposite extreme, where radiative effects make a negligible contribution because there is hardly any radiation, leaving aerodynamic effects to dominate. The F_E perturbation then follows the surface stress, peaking near the crest, with F_H behaving antisymmetrically. In case B both aerodynamic and radiative effects are significant, causing the positions of maximum F_E and F_H to be intermediate between cases A and C. Also, illustrating conclusion (4) above, it is evident that the spatial average (over the entire hill) of

perturbations in available energy is exactly zero, while the spatial averages of the F_E and F_H perturbations are zero apart from a small elevation contribution.

Of the four general conclusions above, the first three concern processes and scaling. These are likely to have general validity beyond the linear theory of RWCH, which is strictly applicable only to low hills without separation, because similar kinds of processes and scaling laws are likely to operate at all hill steepnesses and whether or not the flow separates. In contrast, conclusion (4) is a direct consequence of the use of linear theory, valid only within the domain of low hills without separation. It is likely that nonlinear processes cause spatially averaged flux perturbations to depart from zero when the terrain becomes steep enough, but whether such departures are small or large in practice is a current research question.

3.2 *Nonlinear Effects: an Experiment on Scalar Transfer over Hills*

The consequences of separation and other nonlinear processes in the wake region are difficult to assess theoretically, so we must make best use of experimental techniques. Wind tunnel measurements have a great deal to offer in studying these complex phenomena, because the spatial density of measurements required is very large and because field experiments are subject to natural vagaries in the weather. Of course, field experiments are crucial and have a long history in studies of flow over hills, as already reviewed. However, at this early stage in studies of the wake region, guidance from wind tunnel experiments is appropriate. In the next subsection, the wind tunnel results are extrapolated into the outdoor atmosphere.

We anticipate that local surface-to-air scalar transfers in a hilly region will depend strongly on both surface roughness and hill steepness, because scalar transfer depends strongly upon surface roughness on a flat surface, and because both roughness and steepness have a strong effect on separation, and thence on scalar transfer in the separation region. The profound influence of separation on scalar transfer was shown by wind tunnel measurements in the separated flow behind fences (Sheih *et al.*, 1978).

To explore these effects, we recently carried out a preliminary wind tunnel experiment on the dependence of scalar transfer on steepness and roughness for isolated 2D ridges. This work will be described fully by Finnigan, Raupach and Stanley (in preparation). Figure 10a shows the experimental arrangement, in a small wind tunnel with working section 1.1 m long and

0.4 m in square cross section. A gap-toothed fence, 20 mm high, was placed at the beginning of the working section to stimulate the growth of a thick turbulent boundary layer, which was 150 mm deep at the point where the model hill was placed. The scalar used was passive heat (that is, heat at sufficiently low concentration for the flow to remain thermally neutral). The plane heat source, constructed from heating mats, covered the entire floor of the tunnel including the model hill (Figure 10b). Insulating material below the mats prevented any significant downward heat loss through the tunnel floor, so the surface-to-air heat flux from the mats was uniform and easily determined from the measured electrical power dissipation. The heating mats were covered in turn with a smooth plastic film containing temperature-sensitive liquid crystals which changed colour over the range 25-30 °C from red (cool) to blue (warm), providing a useful semi-quantitative indication of heat transfer patterns.

Two hill shapes (denoted 1, 2) were used, both 2D ridges with the "Witch of Agnesi" profile, Equation (6), truncated and faired smoothly into the surrounding flat surface at $x = \pm 2L$. Both hills had $L = 125$ mm, with heights (H) of 11 mm for the shallow hill 1 and 43 mm for the steep hill 2, after truncation at $x = \pm 2L$ (Figure 10c). Separation in the wake region never occurred on hill 1 and always occurred on hill 2, irrespective of the choice of surface roughness. For each hill shape, three surface roughnesses (a , b and c) were used, consisting of bamboo strips (diameter 2 mm) lying transversely on the liquid-crystal film as shown in Figure 10b, at spacings of 33 mm, 17 mm and 10 mm, respectively. The corresponding momentum roughness lengths (z_0) were respectively 0.045, 0.54 and 0.80 mm, a being the smoothest and c the roughest surface. Simple measurements sufficed to determine the heat transfer: mean wind speeds were measured with a traversed boundary-layer Pitot tube and fixed NPL Pitot-static tube outside the boundary layer; the mean free-stream air temperature with a mercury-in-glass thermometer; mean surface temperatures from the colour of the liquid-crystal film, and more accurately with a Barnes infrared thermometer; and the heat flux from electrical power.

Several checks were first carried out to establish the momentum and heat transfer properties of the three roughnesses on a horizontal surface in the absence of a hill. These showed that the measured roughness lengths for momentum and heat for all surfaces were consistent with available published data.

To examine the perturbations in heat transfer caused by a hill, we use the Stanton number or heat transfer coefficient from the surface to the free stream, $C_H(x)$:

$$C_H(x) = \frac{F_H}{\rho c_p U_\infty [\Theta_0(x) - \Theta_\infty]} \quad (10)$$

where F_H is the surface heat flux, Θ mean temperature, and U mean wind speed; subscripts 0 and ∞ respectively denote conditions at the surface and in the uniform free stream above the boundary layer. From its definition, C_H^{-1} is the aerodynamic resistance for heat transfer between the surface and the free stream, made dimensionless with the free stream velocity. In our experiment, F_H , Θ_∞ and U_∞ were spatially uniform, so variations in C_H^{-1} with streamwise distance x reflected changes in the surface temperature Θ_0 caused by variations in aerodynamic heat transfer.

With no hill present, there is a slow, steady increase of C_H^{-1} with x because of the normal growth of the turbulent boundary layer, and consequent increase of local aerodynamic resistance with x . The model hills cause perturbations about this background situation when placed in the tunnel. In Figure 11, the perturbations in dimensionless resistance C_H^{-1} are shown by plotting $C_H^{-1}(x, \text{hill})/C_H^{-1}(x, \text{no hill})$ against x/L , for the six cases formed from two hills (1 and 2) and three roughnesses (a, b and c). The normalisation of C_H^{-1} with its value at the same x in the absence of the hill is to remove the trend introduced by the growth of the boundary layer and make the hill perturbation more evident. The results indicate quite complex interactions between roughness, topography and heat transfer. Overall features are a rise in the dimensionless aerodynamic resistance in the deceleration region just upwind of the hill, a pronounced fall to low values on the upwind slope of the hill, and very high values in the wake region on the

downwind slope. More detailed scrutiny indicates several points, all of which were experimentally reproducible. Treating first the steep hill 2 (where separation always occurred), the region of increased C_H^{-1} in the separated wake region extended about $2L$ downwind from the crest. Reattachment of the separated flow occurred near this point, inducing a significant fall in resistance (increased transfer) peaking about $2.5L$ from the crest. Surprisingly, this is less pronounced for the intermediate-roughness case (2b) than for cases 2a and 2c. Turning to the shallow hill 1 (where wake-region separation did not occur), there is a much larger sensitivity of the resistance perturbations to roughness than for hill 2, but the largest perturbations occur at the intermediate roughness (1b). At present, these details cannot be explained.

Apart from demonstrating the complexities of roughness-topography interactions, this experiment confirms the limitations of the linear theory discussed above. Since the heat flux was uniform in the wind tunnel experiment, the only effects contributing to perturbations over the hill were aerodynamic. The linear theory indicates approximately symmetric perturbations in this case, peaking at the hill crest (as in Figure 9c, which shows aerodynamically-dominated perturbations in energy partition, closely related to perturbations in C_H^{-1}). In contrast, the experimental results give roughly antisymmetric perturbation patterns. Several reasons can be advanced for this difference: first, separation or near-separation in the wake region, unaccounted for in the linear theory; second, the presence of roughness large enough to occupy a significant fraction of the inner-region depth l ; and third, departure of the approach flow in the experiment from the deep logarithmic boundary layer assumed in the RWCH theory.

3.3 *The Regional Surface Energy Balance in Hilly Terrain*

Finally, we return to the effect of hilly terrain on the regional SEB, by posing this question: if we introduce hills on a flat region (that is, "crinkle" the surface) without otherwise changing its surface properties, how does the regionally averaged SEB change? The linear RWCH theory predicts no change except for small elevation effects, but this prediction cannot be trusted except for very small H/L . How do nonlinear processes alter the picture?

For flat terrain, the influence of heterogeneity on the regional SEB has been the subject of much recent work. Raupach and Finnigan (1994, henceforth RF) tested a "scale-insensitivity"

hypothesis, that the spatially averaged SEB of a patchwork-quilt surface is insensitive to the scale (X) of the patches. This was done with a convective boundary layer (CBL) slab model of the daytime atmosphere, by applying the CBL slab model to patchwork-quilt surfaces with X in both the microscale ($X \lesssim U_m T_*$) and macroscale ($X \gtrsim U_m T_e$) ranges, but with otherwise identical patch area fractions and properties. (Here U_m is the mean wind speed in the CBL, and T_* and T_e the convective and entrainment time scales, respectively; typically, $U_m T_*$ is around 1-5 km and $U_m T_e$ around 100 km). There was very little difference in the diurnal course of the SEB over heterogeneous surfaces with these two very different scales, suggesting that the scale-insensitivity hypothesis is valid at least in a dry (noncloudy) CBL. Other modelling evidence supports this conclusion (see RF).

Here, we apply a similar approach to infer the effect of topography on the regional SEB. The approach is to scale up the above measurements of topographically induced heat transfer perturbations, from hills in a small wind tunnel to hills 1000 times larger in a CBL. The procedure is:

- (1) We assume a vegetated land surface with the simple properties specified in Table 4. This surface is either flat or is "crinkled" with a series of ridges, lateral to the wind, with the shape of hill 2 in the wind tunnel experiment described above, but 1000 times larger; that is, $L = 125$ m and $H = 43$ m. The ridges are $8L$ (1000 m) apart, a separation sufficient to ignore interactions between the ridges. As shown in Figure 12, ridgelines run north-south and the wind blows steadily from the west, so that the sun shines on the lee side in the morning and the windward side in the afternoon.
- (2) We assume that the topographically induced perturbations in scalar (heat and water vapour) transfer over these ridges can be described by a dimensionless resistance $C_H^{-1}(x)$, with the x dependence as measured in the wind tunnel for roughness a . The actual aerodynamic resistance to scalar transfer from the surface to the well-mixed bulk of the CBL is then $r_a(x) = U_m^{-1} C_H^{-1}(x)$, where U_m is the mean CBL wind speed, the counterpart of the free stream wind speed in the wind tunnel experiment (Figure 12). This spatially dependent resistance includes all

the aerodynamic effects occurring around the hill, as it is obtained experimentally from a dynamically similar (though geometrically much smaller) situation.

(3) The surface is assumed to consist of lateral strips, each of which is a patch in a microscale-heterogeneous landscape. The CBL slab model of RF can now be integrated, yielding the daytime course of the CBL temperature and specific humidity, θ_w and Q_w , and the heat and water vapour fluxes from each patch, $F_H(x)$ and $F_E(x)$. These fluxes are calculated using a combination (Penman-Monteith) equation as in RF, with the bulk surface resistance r_s held constant (Table 4), the aerodynamic resistance $r_a(x)$ prescribed for each patch from the wind tunnel results, and the available energy calculated from specified shortwave and longwave irradiances, accounting for slope effects. The assumed atmospheric conditions and computational parameters (identical to RF) are given in Table 5.

Figure 13 shows the x variation of the energy fluxes, at 0900 and 1500, with the bulk surface resistance spatially uniform at 100 s m^{-1} . There are significant perturbations in F_E , F_H and the available energy $F_A = F_E + F_H$, of opposite sign in the morning and afternoon as the sun switches sides of the ridges. The diurnal course of the spatially averaged fluxes $\langle F_E \rangle$ and $\langle F_H \rangle$

is shown in Figure 14, for both a flat landscape and a ridged landscape, and for three choices of the spatially uniform r_s : 10, 100 and 1000 s m^{-1} , corresponding to wet, typical and dry surfaces, respectively. For every choice of r_s , the spatially averaged fluxes are indistinguishable for the flat and the ridged surface. Therefore, despite the significant *local* flux perturbations induced by the topography, the *spatially averaged* flux perturbations are negligible.

This conclusion is in accord with the linear RWCH analysis, but rests on different foundations: there is no assumption of linearity, but instead an application of experimental results to determine the aerodynamic effects on scalar transfer over the ridge in the presence of separation. The present finding is analogous that of RF, who found that the SEB in a flat, heterogeneous, patchwork-quilt landscape is insensitive to patch scale. Both findings are subject

to the limitations of the CBL slab model employed, which ignores mesoscale circulations and clouds.

4. SUMMARY AND CONCLUSIONS

This survey of topographic effects on land-atmosphere interactions began with an acknowledgment that topography influences practically every conceivably relevant process: wind flow (in several ways), radiation, thermodynamics, cloudiness and precipitation, soil properties and ecological properties. We have restricted this discussion to wind flows, mainly over low hills, and to their effects on scalar transfers and especially the SEB.

To discuss the wind and turbulence fields in flows over low hills, we invoked a physical distinction between an inviscid-response outer region, a turbulent-response inner region, and a highly turbulent wake region. We outlined five separate modelling approaches: linear multilayer models, linear mixed spectral finite difference (MSFD) models, nonlinear MSFD models, finite-difference models and large-eddy simulation (LES) models. Through this sequence of approaches there is a decreasing reliance on analytical solutions and an increasing computational demand. We discussed observations by idealising the main experimental results from numerous field and wind tunnel observations of wind flow and turbulence over low hills, for instance in Figures 5 and 7, and then relating these to the flow division into inner, outer and wake regions. The wake region, in particular, has been identified as one where available information "has a distinct signal-to-noise problem" (Taylor, 1988). We have also touched on the effects of ambient stratification, pointing out that the division of the flow into inner and outer regions remains valid in weakly and moderately stable conditions but not in strongly stable conditions (when the Froude number based on hill height is less than 1).

Turning to scalar transfer, we outlined the main findings from linear theory concerning the modification of the SEB over hills. We then addressed nonlinear influences experimentally, using preliminary results from a wind tunnel experiment, translated into the atmosphere with the aid of a CBL slab model. This approach predicts large *local* topographic perturbations in the SEB in hilly regions, caused by a combination of aerodynamic and radiative effects. However, topographic perturbations on *spatially averaged* scalar fluxes prove to be small. This conclusion requires further substantiation because of the limitations of the CBL model used: clouds and

mesoscale circulations are both ignored. If further work shows that topographic perturbations on spatially averaged scalar fluxes are small even in the presence of such complicating (and highly nonlinear) influences, then the result has worthwhile simplifying implications for the parameterisation of sensible and latent heat fluxes in hilly terrain for Global Climate Models (GCMs), where spatially aggregated fluxes over entire regions are required.

This survey has indicated some of the ways in which field experiments, wind tunnel experiments, theory and modelling can elucidate the processes affecting land-air exchanges in hilly terrain. However, much has been treated cursorily or omitted. Among many outstanding issues are: phenomena associated with separation; the regional averaging of the SEB in terrain higher and steeper than we have discussed here; the mutual forcing of meteorological, hydrological and ecological processes in hilly and mountainous terrain; and many more. The subject is complex and very wide-ranging, and present efforts are only the beginning.

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TABLES

Table 1: Some major field experiments on flow over low hills, showing hill height, H ; length, L ; roughness length, z_0 ; inner layer depth, l , from Equation (2); and observed height of maximum speedup, z_{max}

Hill	Reference	H (m)	L (m)	z_0 (m)	l (m)	z_{max} (m)
Black Mountain	Bradley (1980)	170	275	1.14	28	27
Bungendore Ridge	Bradley (1983)	7.5	75	0.005	3.4	-
Kettles Hill	Taylor <i>et al.</i> (1983)	100	520	0.01	22	5
Nyland Hill	Mason (1986)	70	100	0.04	6.3	3
Askervein	Taylor and Teunissen (1987); Mickle <i>et al.</i> (1988); Salmon <i>et al.</i> (1988)	116	250	0.03	13	4
Cooper's Ridge	Coppin <i>et al.</i> (1994)	115	400	0.05	21	5

Table 2: Conditions for separation in the wake region, for nonstratified hill flow.
 Separation is likely when maximum hill slope on the lee side, θ , exceeds
 values shown

Hill shape	Hill roughness	Lee slope θ
2D ridge normal to mean wind	smooth	inducing separation
	20°	
3D axisymmetric hill	very rough	10°
	smooth	30°
	very rough	20°

Table 3: Background energy fluxes for hill calculations in Figure 9

Case	F_{A1} (W m ⁻²)	F_{E1} (W m ⁻²)	F_{H1} (W m ⁻²)	$\alpha_1 = F_{E1}/F_{A1}$
A	500	347	153	0.694
B	250	239	11	0.956
C	10	136	-126	13.63

Table 4: Surface properties assumed for calculations of SEB over repeated ridges using CBL slab model and wind tunnel specification of $r_a(x)$

albedo α_s	0.2
emissivity ϵ_s	1.0
roughness length z_0 (m)	0.045
bulk surface resistance r_s (s m^{-1})	1, 100, 10^6

Table 5: Meteorological conditions and computational parameters for calculations of SEB over repeated ridges using CBL slab model and wind tunnel specification of $r_a(x)$

Downward shortwave irradiance F_{s_i} on horizontal surface	$F_{s_{\max}} \sin(2\pi t/T_{\text{day}})$ with $F_{s_{\max}} = 800 \text{ W m}^{-2}$, $T_{\text{day}} = 86400 \text{ s}$
Downward longwave irradiance F_{L_i} on horizontal surface	Swinbank (1963) formula: $F_{L_i} \text{ (W m}^{-2}\text{)} = 5.31 \times 10^{-14} T^6 \text{ (T in deg K)}$
Mean wind speed in CBL, U_m	5 m s^{-1}
Conditions above CBL	$\theta_s(z) = 15 + 0.005z \text{ (C)}$ $Q_s(z) = 0 \text{ (g kg}^{-1}\text{)}$
Initial (0600) conditions	$\theta(0) = 10 \text{ C}$ $Q(0) = Q_{\text{sat}}$ $Z_i(0) = 50 \text{ m}$
Integration period Integration time step	12 hr (0600 to 1800 local time) 15 min

FIGURE CAPTIONS

Figure 1: Typical topographically induced perturbations in net irradiance, near-surface air temperature, saturation deficit, wind speed, surface resistance, for nonstratified flow over a series of ridges normal to the wind, with the sun shining on upslopes.

Figure 2: Flow regions and length scales for nonstratified flow over a low hill.

Figure 3: (a) Schematic diagrams of typical perturbations in (a) pressure $P(x,z)$, and (b) shear stress $\tau(x,z)$, for nonstratified flow over a low hill.

Figure 4: Main features of the velocity field for nonstratified flow over a 2D ridge normal to mean wind, and a 3D axisymmetric hill.

Figure 5: Typical observed profiles of mean velocity U for nonstratified flow over a low hill.

Figure 6: Comparison of measured speedup at Cooper's Ridge, New South Wales, with prediction of HLR theory, Equation (5), for both the actual hill shape and a fitted "Witch of Agnesi" hill shape. After Coppin *et al.* (1994).

Figure 7: Typical observed profiles of velocity standard deviations σ_u , σ_v , and σ_w , and Reynolds shear stress $\tau = -\overline{uw}$, for nonstratified flow over a low hill. All quantities are normalised by background values at the same height above ground (subscript b).

Figure 8: Stratified flow over a 2D ridge normal to the wind: inner-layer speedup (normalised with $U_b(h_m)$) as a function of Froude number F_L .

Figure 9: Predictions of RWCH theory for perturbations in surface latent heat flux F_E , sensible heat flux F_H and available energy flux F_A . See text and RWCH for details.

Figure 10: Preliminary wind tunnel experiment on heat transfer over 2D ridges of varying steepness and roughness: (a) tunnel arrangement; (b) construction of heated rough surfaces; (c) hill shapes.

Figure 11: Streamwise variation of dimensionless resistance $C_H^{-1}(x, \text{hill})/C_H^{-1}(x, \text{no hill})$ observed in wind tunnel, for two hills (1, 2) and three roughnesses (a, b, c).

Figure 12: The situation assumed for calculations of effect of ridges on regional SEB, using wind tunnel data on $r_a(x)$ and CBL slab model.

Figure 13: Predictions of CBL slab model with wind tunnel $r_a(x)$ for variation with x of F_A, F_E and F_H , at 0900 and 1500, with $r_s = 100 \text{ s m}^{-1}$.

Figure 14: Predictions of CBL slab model with wind tunnel $r_a(x)$ for diurnal course of (a) $\langle F_E \rangle$ and (b) $\langle F_H \rangle$, for flat and ridged landscapes with $r_s = 10, 100$ and 1000 s m^{-1} .

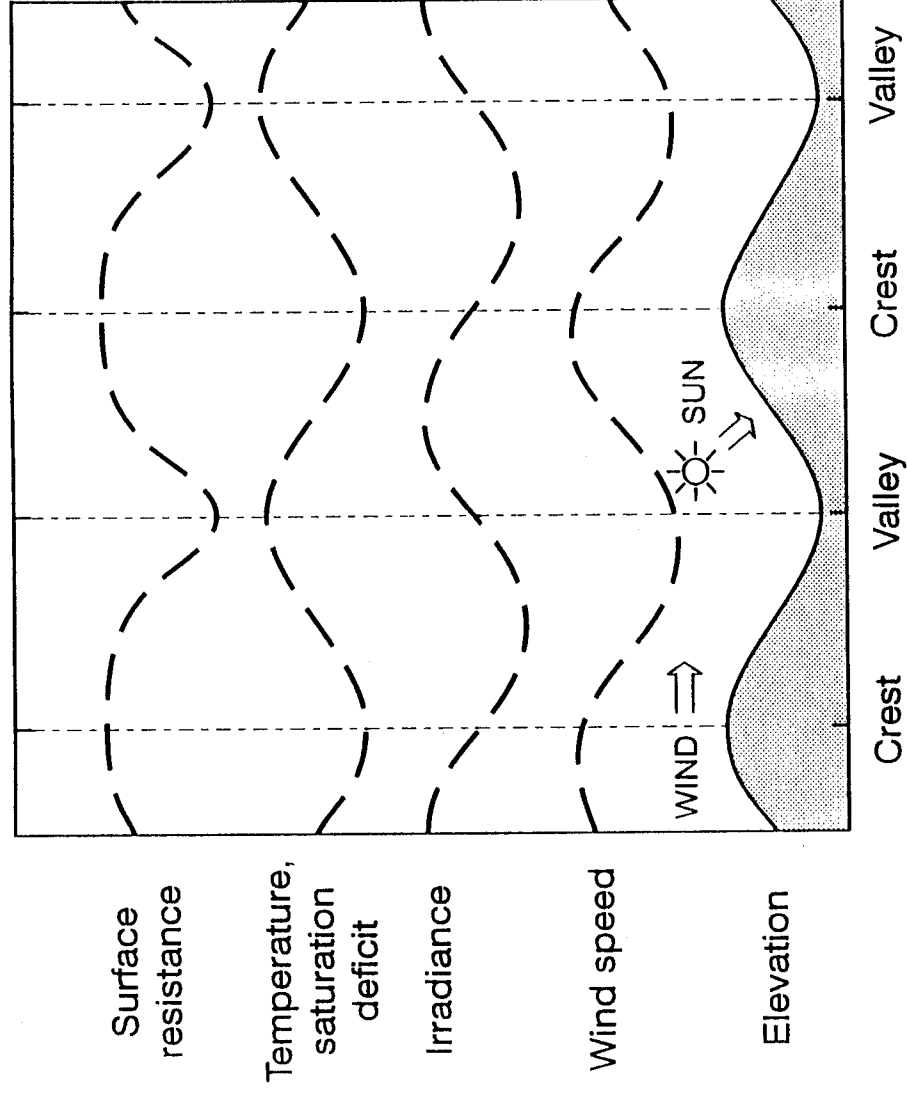


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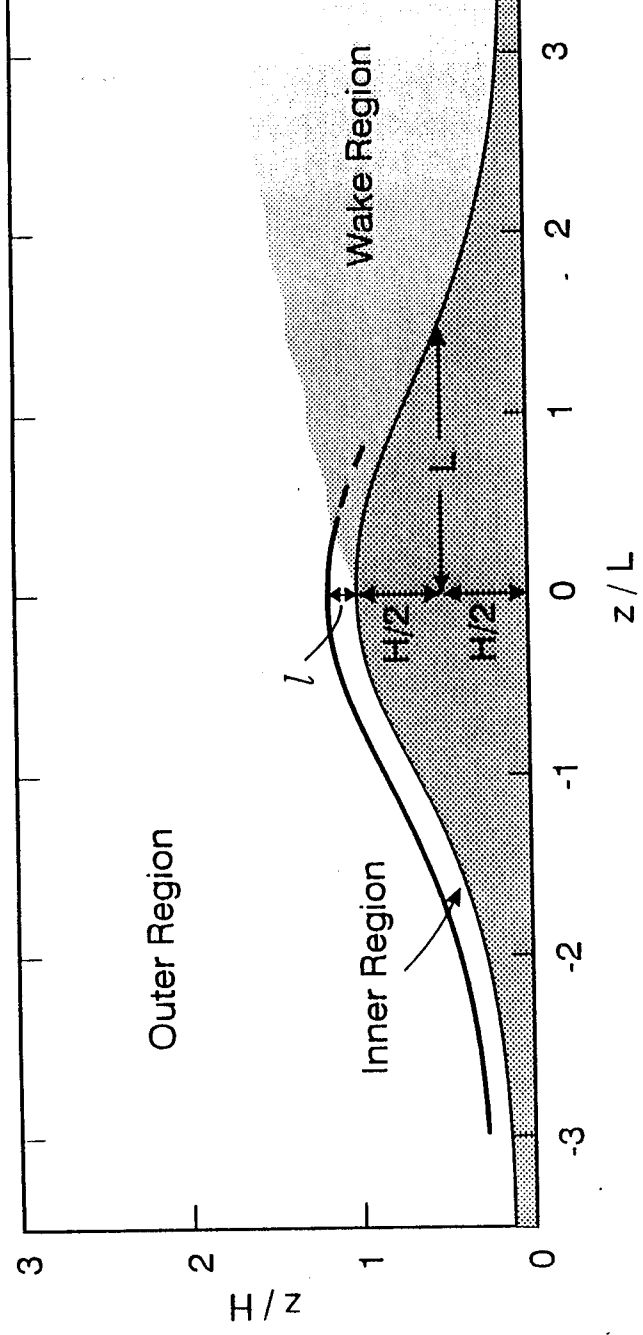
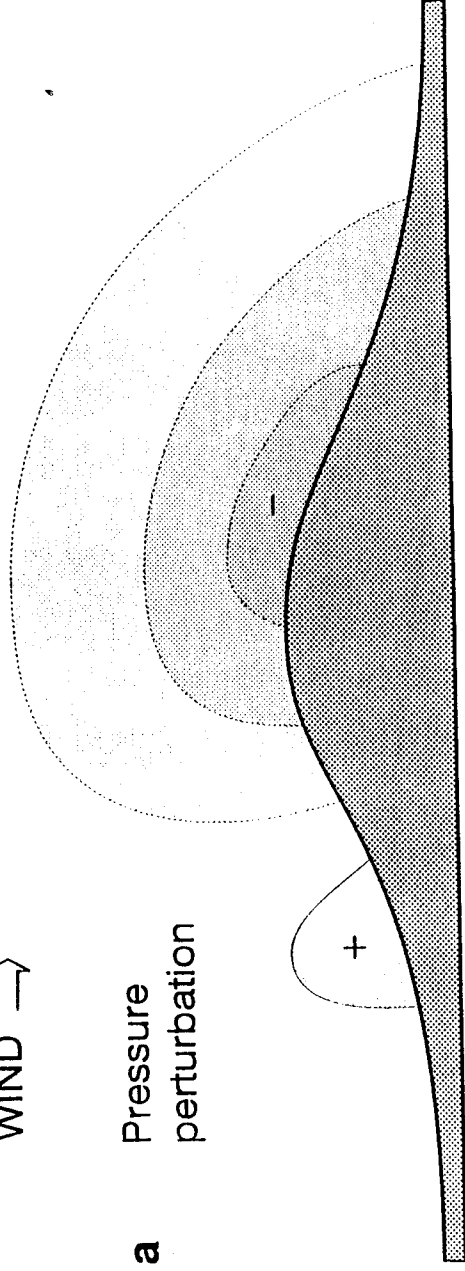


Figure 2: Flow regions and length scales for nonstratified flow over a low hill.

WIND $\Rightarrow$

a Pressure
perturbation



b Stress
perturbation

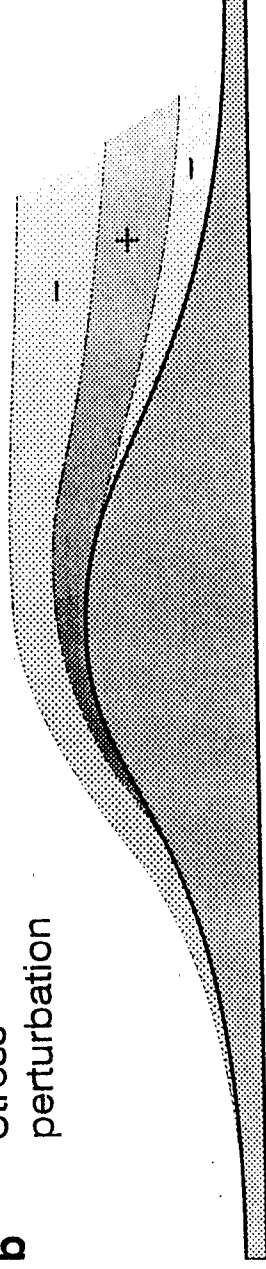


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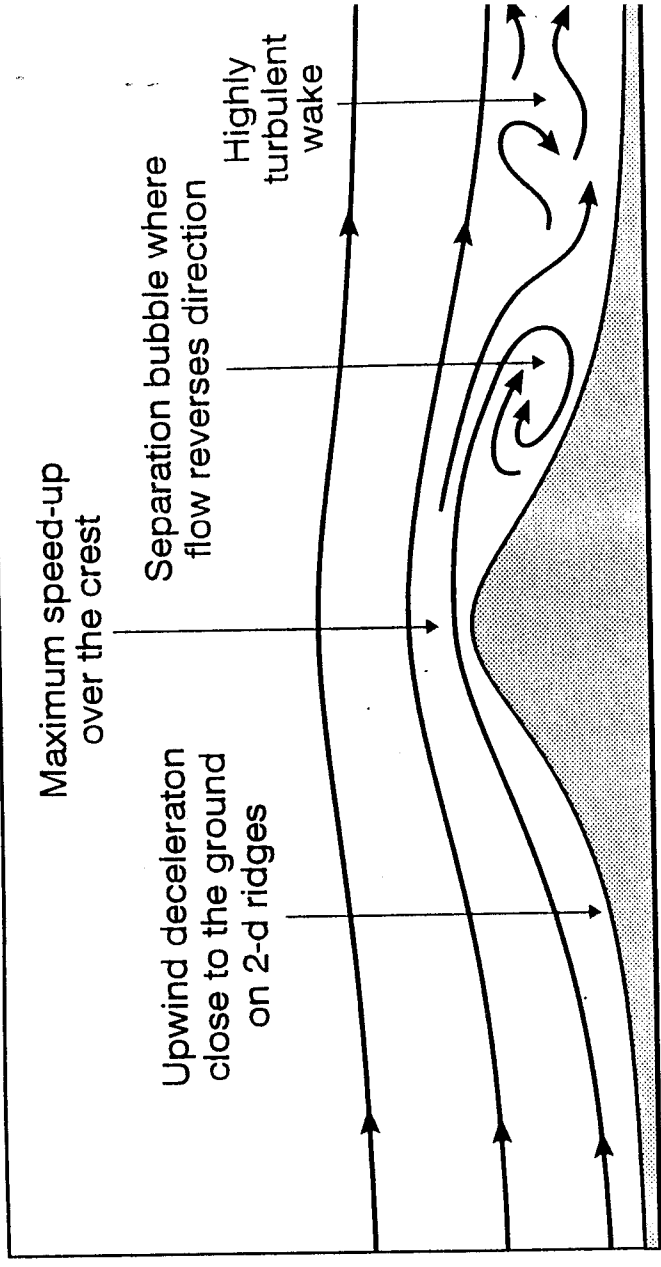


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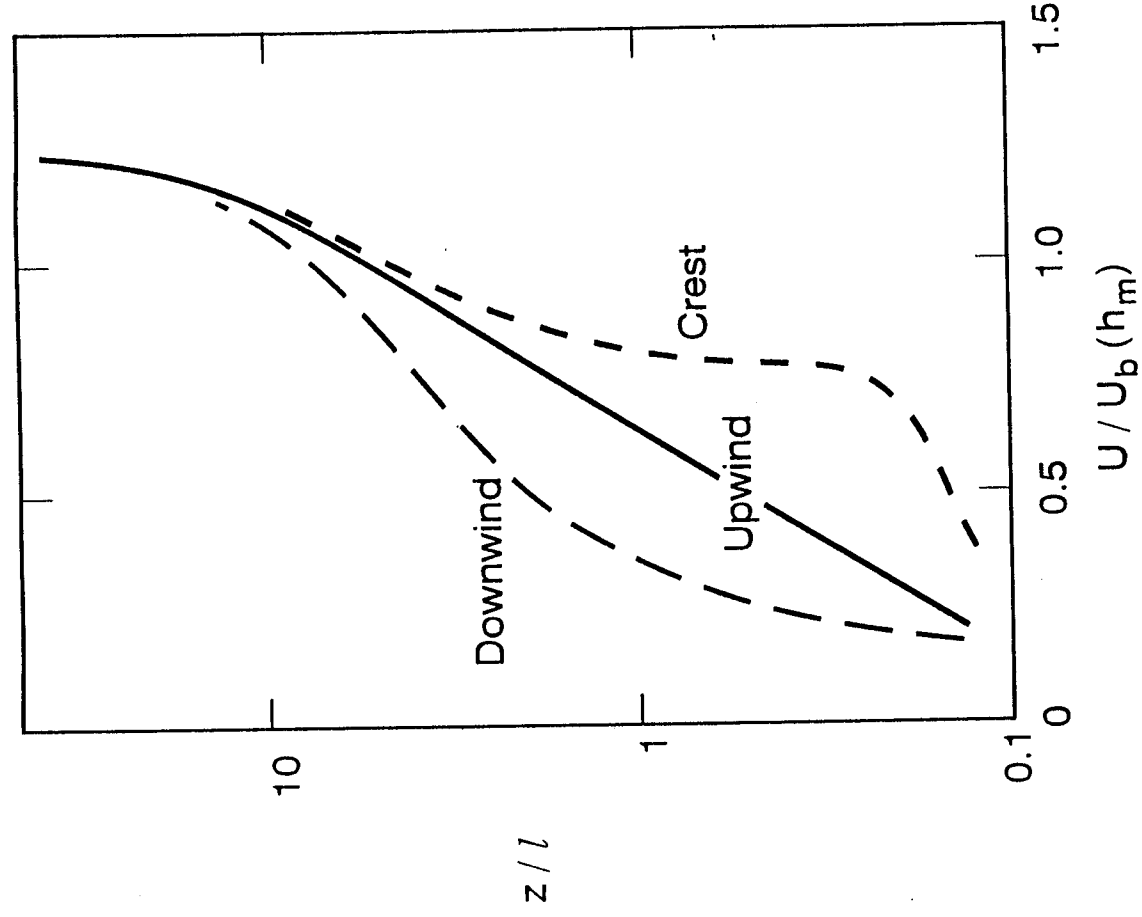


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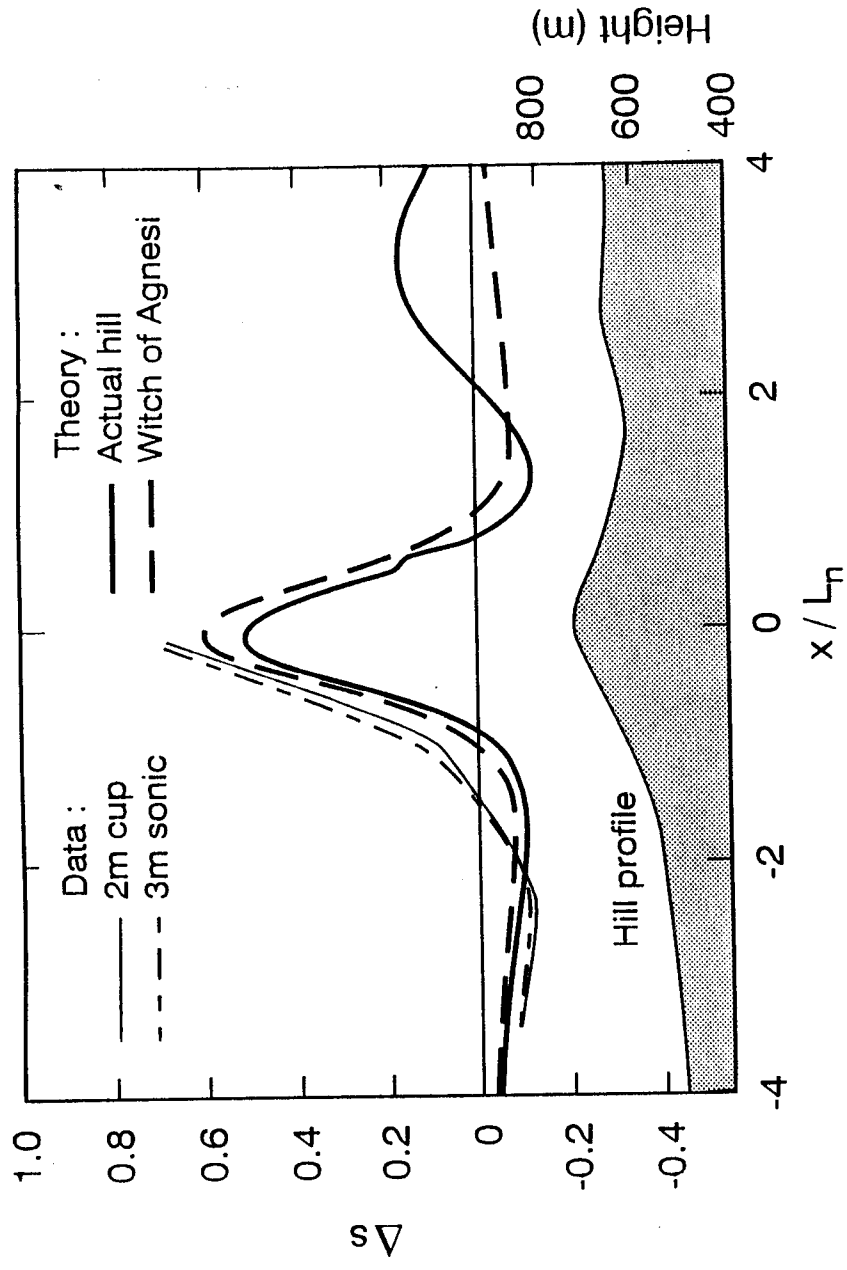


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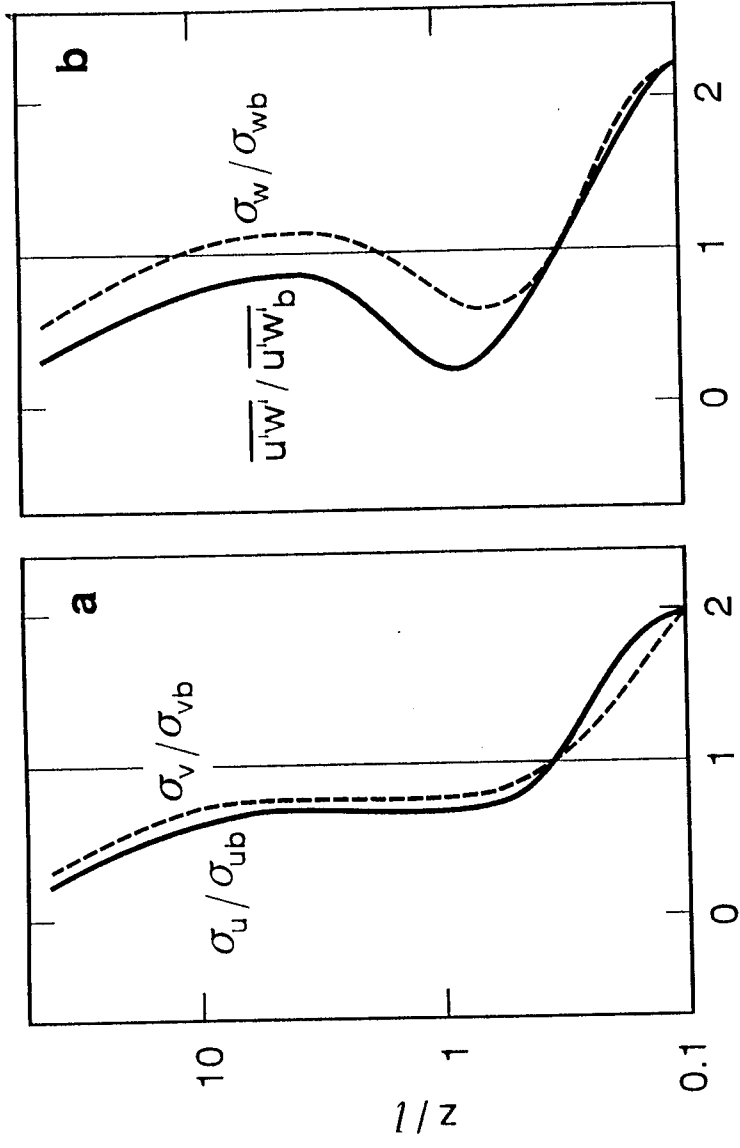


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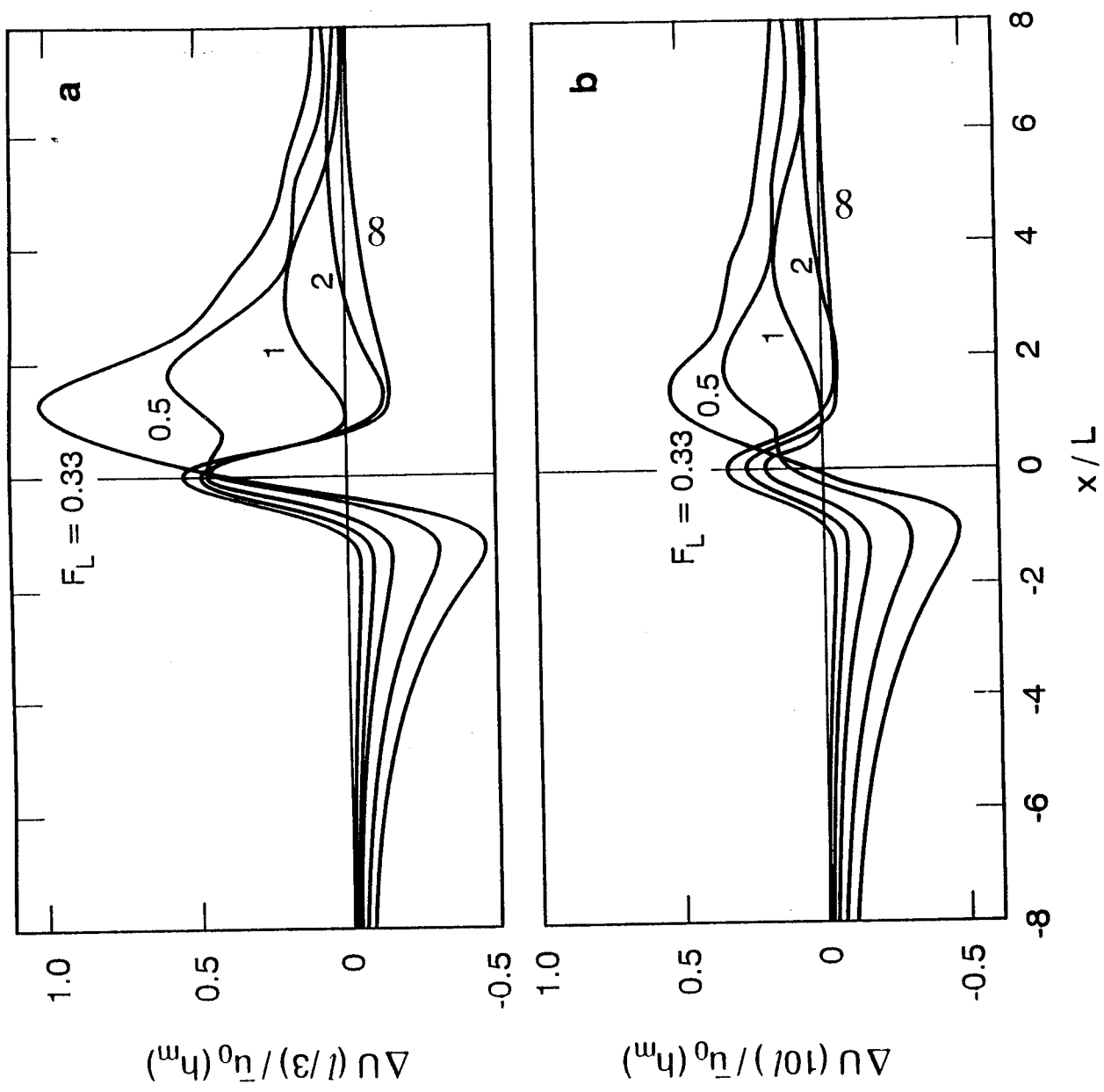


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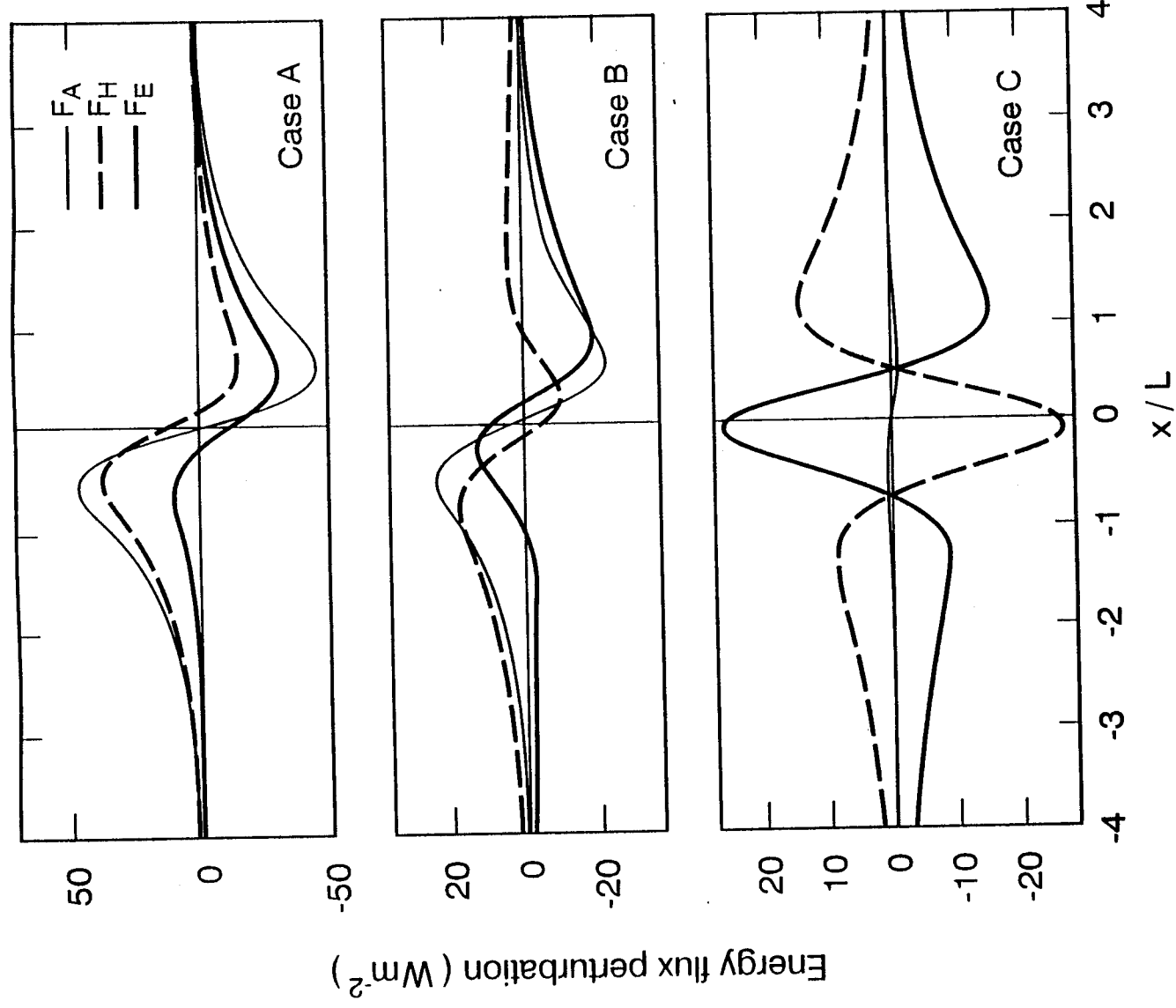


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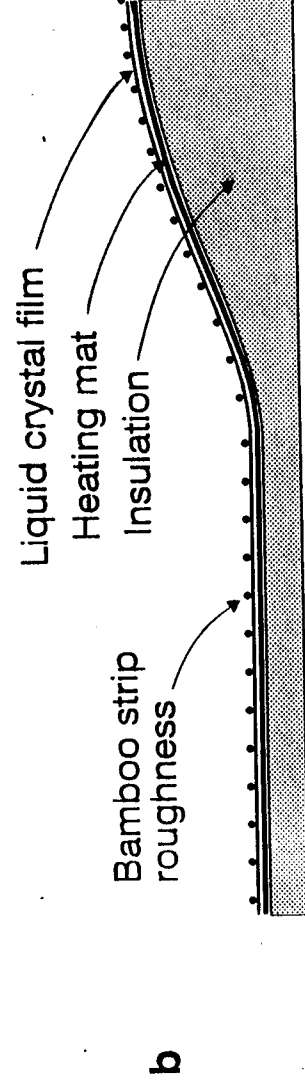
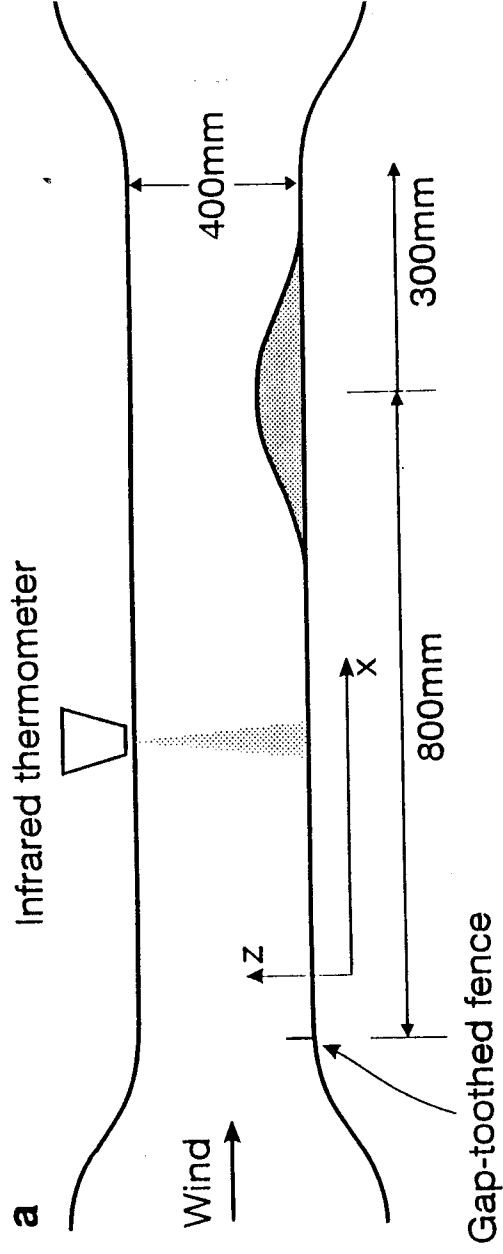


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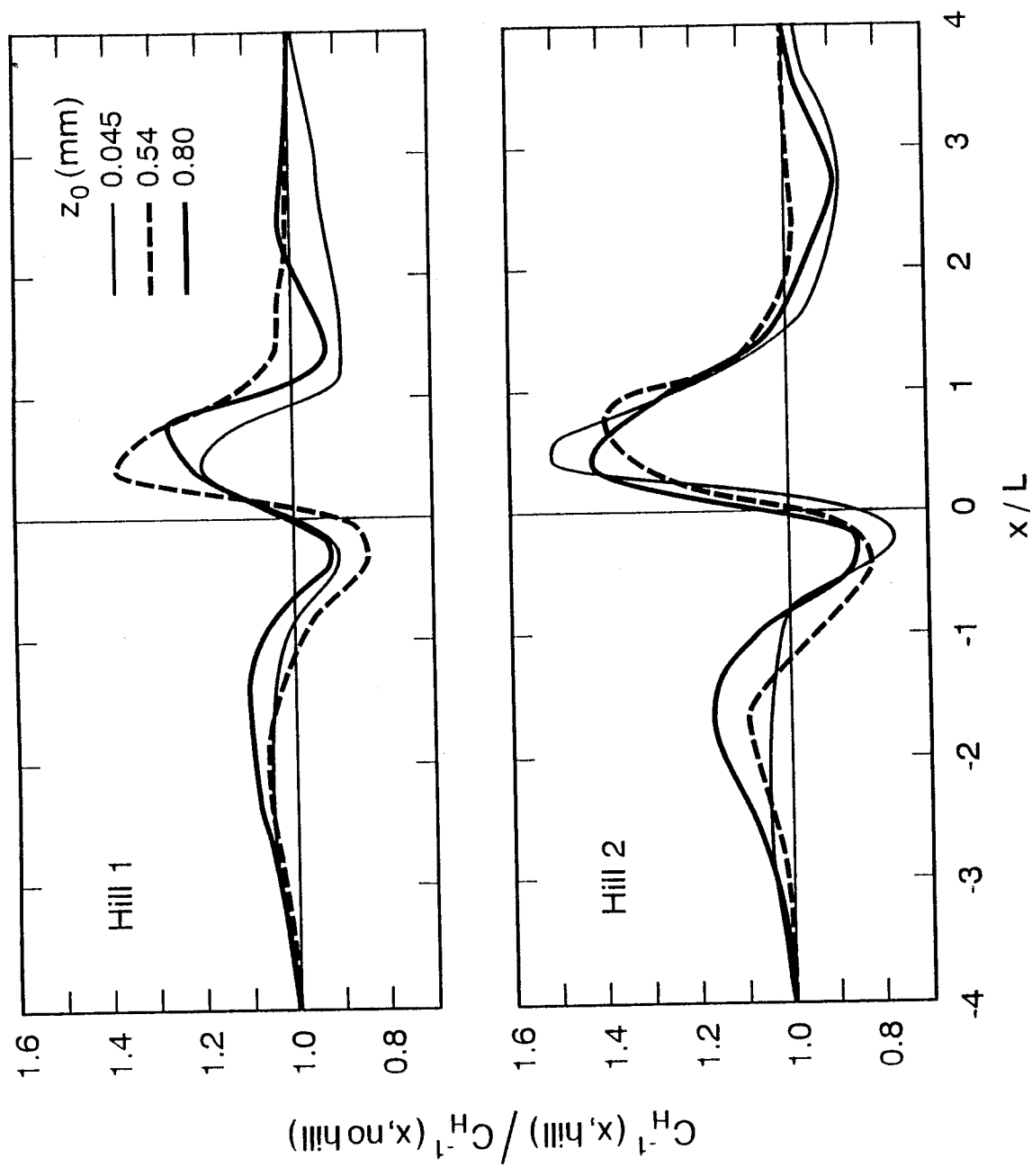


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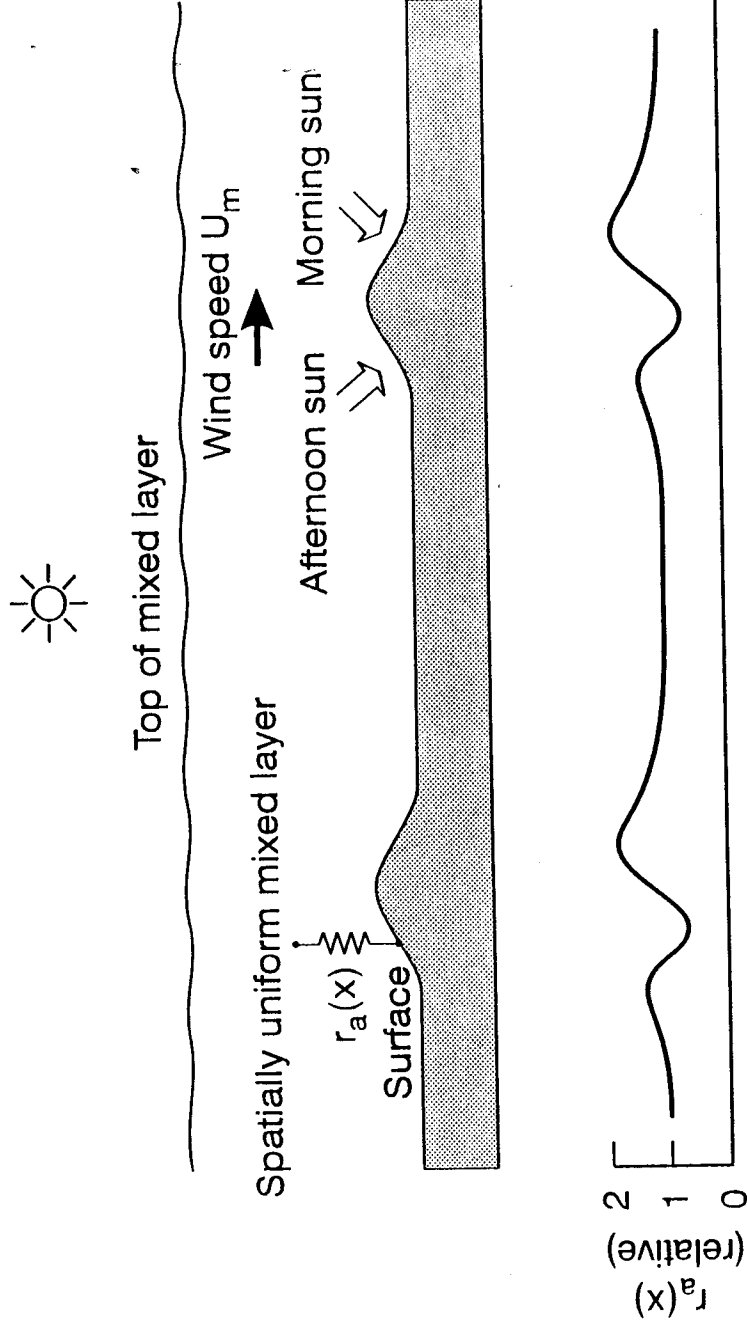


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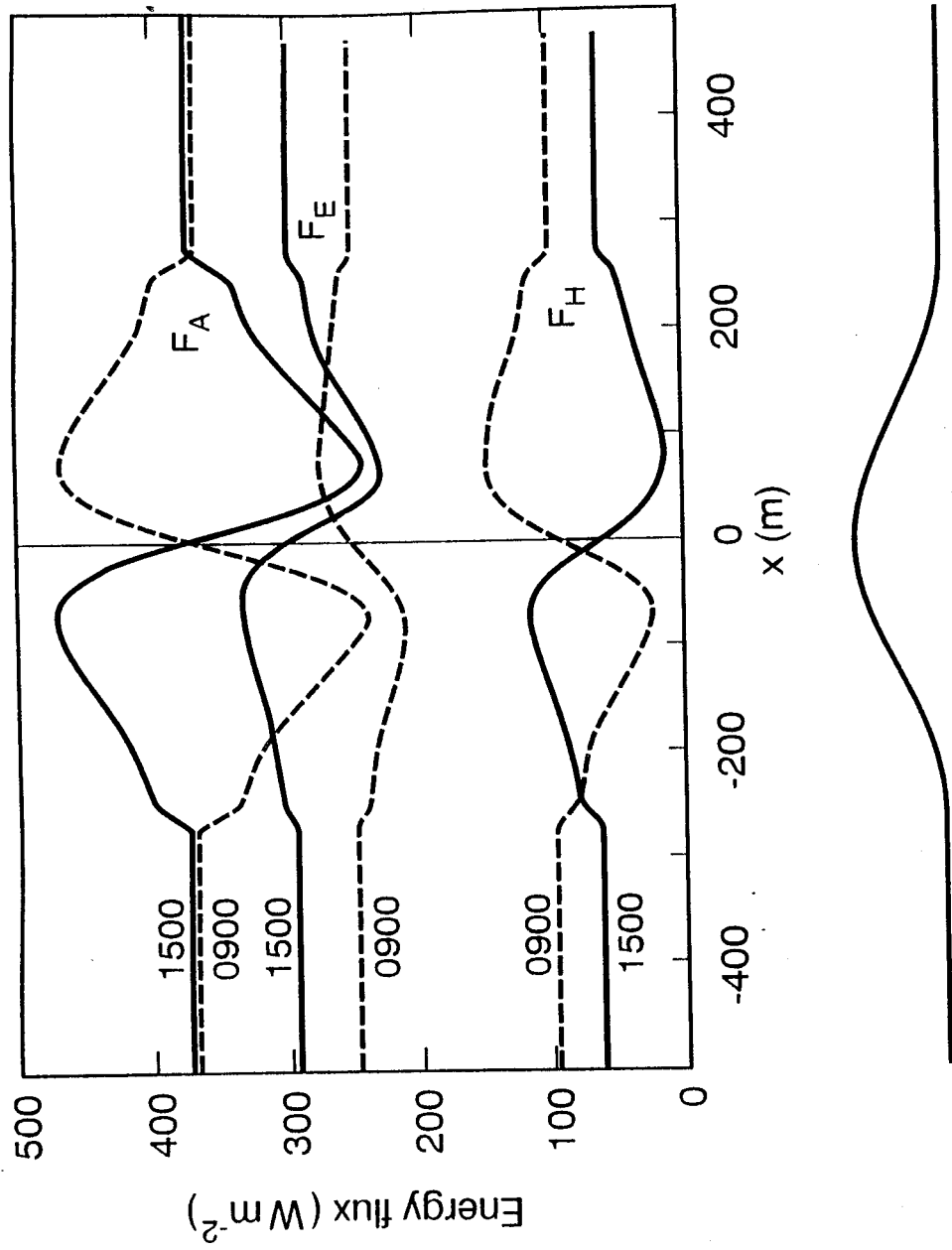


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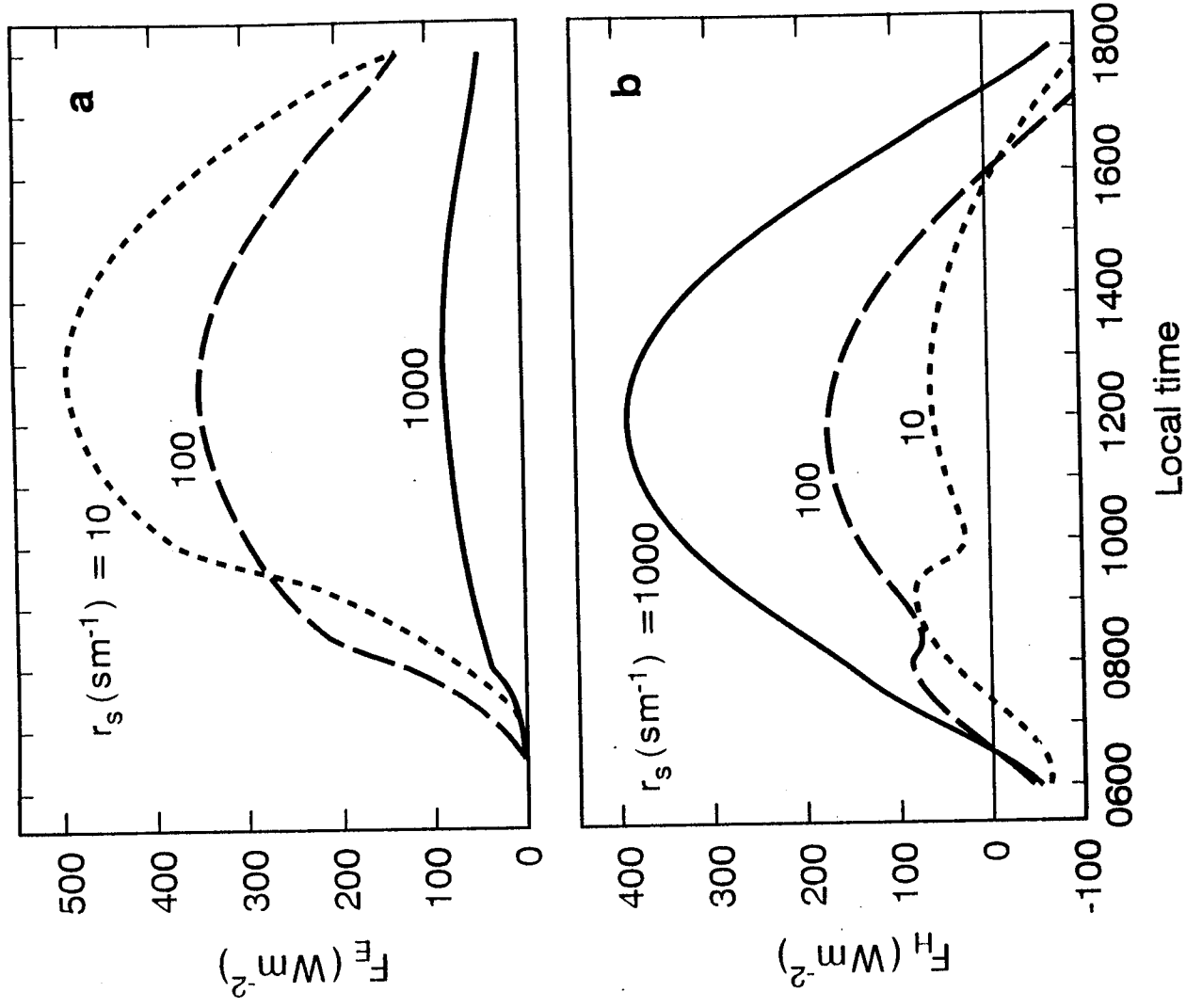


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The impact of using area-averaged land surface properties - topography, vegetation condition, soil wetness - in calculations of intermediate scale (~ 10 km²) surface-atmosphere heat and moisture fluxes.

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Abstract

It is commonly assumed that biophysically-based soil-vegetation-atmosphere transfer (SVAT) models are scale-invariant with respect to the initial boundary conditions of topography, vegetation condition and soil moisture. In practice, SVAT models that have been developed and tested at the local scale (a few meters) are applied almost unmodified within general circulation models of the atmosphere (GCMs), which have grid areas of (50 to 500 km)². This study, which draws much of its substantive material from the paper of Sellers et al. (1992b; in prep.) explores the validity of doing this. The work makes use of the FIFE-89 data set which was collected over a 2 x 15 km grassland area in Kansas which was characterized by high variability in soil moisture and vegetation condition during the late growing season of 1989. The area also has moderate topography.

The 2 x 15 km testbed area was divided into 68 x 501 pixels of 30m x 30m spatial resolution, each of which could be assigned topographic, vegetation condition and soil moisture parameters from the satellite and in situ observations gathered in FIFE-89. One or more of these surface fields was area-averaged in a series of simulation runs to determine the impact of using large-area means of these initial/boundary conditions on the area-integrated (aggregated) surface fluxes. The results of the study can be summarized as follows:

- (i) Analyses and some of the simulations indicated that the relationships describing the effects of moderate topography on the surface radiation budget are near-linear and thus largely scale-invariant. The relationships linking the simple ratio vegetation index (SR), the canopy conductance parameter (∇F) and the canopy transpiration flux are also near-linear and similarly scale-invariant to first order, see also Sellers et al (1992b, c). Because of this, it appears that simple area-averaging operations can be applied to these fields with relatively little impact on the calculated surface heat flux.
- (ii) The relationships linking surface and root-zone soil wetness to the soil surface and canopy transpiration rates are non-linear.

However, simulation results and observations indicate that soil moisture variability decreases significantly as an area dries out, which partially cancels out the effects of these non-linear functions.

In conclusion, it appears that simple averages of topographic slope and vegetation parameters can be used to calculate surface energy and heat fluxes over a wide range of spatial scales, from a few meters up to many kilometers. Although the relationships between soil moisture and evapotranspiration are non-linear for intermediate soil wetnesses, the dynamics of soil drying act to progressively reduce soil moisture variability and thus the impacts of these nonlinearities on the area-averaged surface fluxes. These findings indicate that we can use mean values of topography, vegetation condition and soil moisture to calculate the surface-atmosphere fluxes of energy, heat and moisture at larger length scales, to within an acceptable accuracy for climate modeling work.

1. Introduction

The land surface-atmosphere models currently used within general circulation models of the atmosphere (GCMs) are largely based on formulations of processes as observed or conceptualized at relatively small spatial scales, i.e. a few meters. In applying these models within GCMs, there is an implicit assumption of scale-invariance in almost all of the model components. While there are some notable exceptions to this generalization; for example, the treatment of non-spatially uniform precipitation in the Simple Biosphere model as reported by Sato et al. (1989) and of soil wetness spatial distributions by Wetzel and Chang (1987), by and large, most modelers simply drive unaltered versions of their soil- vegetation-atmosphere-transfer (SVAT) models at GCM grid scales using assumed area-averaged lower boundary conditions (vegetation properties, topography, soil moisture) and atmospheric and radiative forcings. The use of increasingly sophisticated SVATs in GCMs, see for example Dickinson (1984), Sellers et al. (1986; 1995a, b), Bonan (1994), has almost always resulted in improved model performance and the simulation of more realistic continental climatologies, see for example Sato et al. (1989), Randall et al. (1995), Dickinson and Henderson-Sellers (1988), Nobre et al. (1991). However, in parallel with this trend of increasing model sophistication has been a growing concern that the scaling assumptions inherent in them may not be valid and that specifying parameters for these models will become increasingly difficult and at some level, meaningless. These two issues were aired in many discussions in the early and mid-1980's and were the motivation for the International Satellite Land Surface Climatology Project (ISLSCP) program of field experiments which have the goal of:

1. Upscale integration of models: Experiments were designed to test the soil-plant-atmosphere models developed by biometeorologists for small-scale applications (millimeters to meters) and to develop methods to apply them at the larger scales (kilometers) appropriate to atmospheric models and satellite remote sensing.
2. Application of satellite remote sensing: The experiments are tasked with exploring methods for using satellite data to quantify important biophysical states and rates for model input, with the ultimate aim of providing a means of initializing and validating SVATs on regional and global scales.

This paper reviews the results of some recent work carried out with a biophysical SVAT, the Simple Biosphere model (SiB) of Sellers et al. (1986) which used data from the First ISLSCP Field Experiment (FIFE) to explore the impact of area-averaging initial/boundary conditions of topography, vegetation condition and soil moisture on the calculate fields of radiation, heat and water vapor, see Sellers et al. (1992a). In the process, methods of defining surface properties from satellite data were developed and tested.

2. Theoretical Consideration in 'Upscaling' Soil-Vegetation-Atmosphere Transfer (SVAT) Models

2.1 The Simple Biosphere Model (SiB)

The current generation of biophysically realistic SVATs is well-represented by the BATS model of Dickinson (1984), SiB of Sellers et al. (1986), and LSM of Bonan (1994). In many respects, these models are very similar in their underlying approach and assumptions. All three have explicit representations of the vegetation canopy overlying multi-layer soil models.

SiB was modified in Sellers et al. (1992b) to run independently of a GCM with the required atmospheric forcings provided by near-surface observations of meteorological conditions and downwelling radiation fluxes. In Sellers et al. (1992b), data from the FIFE network of automatic meteorological stations (AMS) were used to provide these upper boundary conditions.

Figure 1 shows the structure of SiB: it can be seen that the transfer of soil water into the atmosphere is regulated by a canopy resistance (r_c), a soil surface resistance, (r_{surf}) and a network of aerodynamic resistances; r_a , r_b , r_d . In many respects, the treatment of the energy and heat balance by SiB can be thought of as an extension of the Penman-Monteith equation, see Monteith (1973). Like just about all SVATs in use today, SiB has been demonstrated to provide realistic local-scale simulations of the surface sensible and latent heat fluxes for many different vegetation types, see for example Sato et al. (1989), Sellers et al. (1992b).

2.2 Scale invariance in SVATs

The issue of scale-invariance with regard to SVAT performance is the extent to which the calculated large-domain land-atmosphere fluxes of heat and water are impacted by the use of area-averaged surface boundary conditions; specifically, the fields of topography, vegetation condition and soil moisture. We can address this issue in abstract terms by defining the dependence of a surface flux, Q , which in practice would be sensible heat flux H , or latent heat flux LE ; on some surface quantities x_i ; which in practice could be three variables mentioned above.

Thus, if we have a model for Q as a function of x_i which we believe works well at the local scale, (see Figure 2), we may write:

$$Q = f(x_1, x_2, x_3, \dots) \quad (1)$$

As exact calculation of Q over a large domain of area A , i.e. a representative area-averaged value of Q , $\langle Q \rangle$, may be obtained from:

$$\langle Q \rangle_I = \frac{1}{A} \oint_0^A f(x_1, x_2, x_3) da \quad (2)$$

$\langle Q \rangle_I$ = estimate of area-averaged value of surface flux provided by numerical integration.

The angle-brackets ' $\langle \rangle$ ' denote 'area-average'. Figure 2 (left hand side) shows how a numerical calculation may be done to yield $\langle Q \rangle_I$ by dividing up the domain A into many small-scale segments for each of which (1) can be assumed to be valid. In this approach, the fields of x_i are discretized over the domain so that we can calculate Q for each contributing pixel. For practical reasons, this is not done in GCMs. Instead, it is assumed that

$$\langle Q \rangle_M \approx f(\langle x_1 \rangle, \langle x_2 \rangle, \langle x_3 \rangle) \quad (3)$$

$\langle Q \rangle_M$ = estimate of area-averaged value of Q provided by taking mean values of x_i over the domain.

Figure 2 (right hand side) shows how this use of (3) reduces the tedious numerical integration of (2) to a single calculation but, as $f(x_i)$ becomes more non-linear, the difference between the 'true' value of $\langle Q \rangle_I$ as given by (2), and the area-averaged estimate $\langle Q \rangle_M$ provided by (3) will increase.

Under what conditions will the two estimates of $\langle Q \rangle$ be approximately equal? Figure 3a shows how for linear functions of $f(x_i)$, the two methods will yield the same result. This we can write:

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_I, \text{ as } \frac{\partial^2 f(x_i)}{\partial x_i^2} \rightarrow 0. \quad (4a)$$

Alternatively, if the spatial variability of the x_i field over the domain is very low, the two methods will agree even if the functions are non-linear, see Figure 3b.

Thus:

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_I, \text{ as } \frac{\sigma_{x_i}}{\langle x_i \rangle} \rightarrow 0. \quad (4b)$$

$$\sigma_{x_i} = \text{standard deviation of } x_i \text{ inside the domain.}$$

It is more likely in nature for $f(x_i)$ to be somewhat non-linear and the fields of x_i to show some degree of spatial variability so it is the combination of these factors that is important. Thus we may write a general criterion which combines (4a) and (4b):

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_1, \text{ as } \frac{\partial^2 f(x_i)}{\partial x_i^2} \cdot \frac{\sigma_{x_i}}{\langle x_i \rangle} \rightarrow 0. \quad (4c)$$

How does this analysis transfer to the actual functioning of a typical biophysical SVAT?

2.3 Theoretical effects of variability in surface quantities on the surface energy balance

The surface quantities considered in this study are topography, vegetation condition and soil moisture. The effects of spatial variation in each of these on the surface energy balance is briefly reviewed below:

- (i) Topography: Topographic variations have a direct effect on the absorption and reflectances/emission of radiation by the surface. In addition, topography determines how soil moisture is redistributed when the soil is wet and thereby to a large extent determines an area's streamflow response. The work in this paper addresses only the effects of topography on the radiation balance; for the examples considered, the role of interflow and runoff is small. (Discussion at the end of this paper revisits this issue).

The amount of shortwave (S_n) and Longwave (L_n) radiation absorbed by a pixel within the domain depends on slope and aspect, see Figure 4. In this study, we adapted the SiB radiation submodel to take account of illumination geometry. The modification is simple and does not take into account the effects of topographic shadowing by other pixel elements in the scene.

(5a)

$$\begin{aligned}
S_n &= [S_{\downarrow b,v}(1-a_{b,v}) + S_{\downarrow b,n}(1-a_{b,n})]\beta_b \\
&+ [S_{\downarrow d,v}(1-a_{d,v}) + S_{\downarrow d,n}(1-a_{d,n})]\beta_d \\
&+ [S_{\downarrow b,v}a_{b,v} + S_{\downarrow d,v}a_{d,v}](1-a_{d,v})\beta_r \\
&+ [S_{\downarrow b,n}a_{b,n} + S_{\downarrow d,n}a_{d,n}](1-a_{d,n})\beta_r
\end{aligned}$$

S_n = net shortwave radiation absorbed by a pixel,
 Wm^{-2}

$S_{\downarrow \mu,\lambda}$ = normal downward shortwave flux for angular
component ' μ ' and waveband ' λ '

μ = b (solar beam); d (diffuse)
 λ = v (visible); n (near-infrared)
 $a_{\mu,\lambda}$ = surface spectral reflectance

(5b)

$$\beta_b = 1 + \tan V \tan Z \cos(\alpha_v - \alpha_z)$$

(5c)

$$\beta_d = \left(\frac{\pi - V}{\pi} \right) \frac{1}{\cos V}$$

(5d)

$$\beta_r = \frac{V}{\pi} \frac{1}{\cos V}$$

β_b = Direct beam radiation correction factor for
slope and aspect
 β_d = Fraction of sky dome visible to pixel
 β_r = Fraction of sky dome below the horizon for a
sloping pixel.
 Z = solar zenith angle
 V = slope angle
 α_v, α_z = slope azimuth (aspect); and solar azimuth,
respectively

The first term in (5) deals with the absorption of the direct beam radiation, the second term deals with the interception of diffuse sky radiation and the third and fourth terms cover the absorption of radiation reflected from the surrounding

landscape which would be exposed to a sloping pixel from the horizon downwards. Note that we assume that the landscape has the same reflectance properties as the subject pixel and that there is no shadowing of the subject pixel by distant topography. (A full-up radiative transfer treatment of these issues would be very complex and computationally expensive).

The direct beam surface reflectances, ab_v and ab_n , are a function of the solar beam illumination angle, see Sellers (1985). A correction is made in this treatment to make these reflectances a function of the solar zenith angle as referenced to the slope normal. This 'effective' solar zenith angle, Z' , which is used to calculate ab_n , is given by:

$$\cos Z' = \cos Z \cos V + \sin Z \sin V \cos (\alpha_Z - \alpha_V) \quad (5e)$$

The treatment of longwave radiation exchanges is similar to that for shortwave diffuse radiation.

$$L_n = L \downarrow \beta_d + \epsilon \sigma (f_c T_c^4 + (1-f_c) T_g^4) \left(\beta_r - \frac{1}{\cos V} \right) \quad (5f)$$

L_n	=	net longwave radiation absorbed by the pixel,
		$W m^{-2}$
$L \downarrow$	=	normal longwave flux (diffuse; thermal),
		$W m^{-2}$
σ	=	Stefan-Boltzman constant, $W m^{-2} K^{-4}$
f_c	=	fraction of ground covered by vegetation
T_c, T_g	=	Canopy, ground surface temperature, K

The effects of topography on the radiation budget are encapsulated in the factors β_b , β_d and β_r , see (5b, c, d). Generally speaking, deviations from a flat surface will lead to an increase in the absorption and emission of longwave radiation with a probably small net gain for the surface as the fraction of cold sky viewed by the surface decreases. The impact of topography on the interception of diffuse

radiation is fairly small as the terms multiplied by β_d and β_r tend to decrease slowly from the solution for a level surface as V increases. The absorption of direction beam radiation is almost linearly related to V through $\tan V$ in β_b .

An inspection of (5b, c, d) shows that for moderate values of V , say 25° or less, the impact of slope angle on absorbed radiation is nearly-linearly related to V . (The $\cos V$ terms are relatively weak for values of V of 0 to 25°). Because of this, we might expect some small changes in net radiation over a domain due to increased topography but overall, the effects are partially self-canceling over a day (south slope gains roughly cancel out north slope losses for direct beam radiation, for example) and should be well-behaved owing to the near-linearity of the dependence of S_n on V . The formulation of (5a) thus roughly conforms to the condition specified in (4a), and so we anticipate that simple area-averaging of topography will have little impact on the net radiation and the total heat fluxes, $\langle Q \rangle$, for a domain with no significant net slope.

(ii) Vegetation Condition: In SiB, the canopy transpiration rate, λE_{ct} is calculated from a simple resistance model, see Figure 1.

$$\lambda E_{ct} = \frac{(e^*(T_c) - e_a)}{r_c + r_b} \frac{\rho c_p}{\gamma} \quad (6a)$$

$e^*(T_c)$ = saturated vapor pressure inside canopy foliage, Pa

e_a = vapor pressure in canopy air space, Pa

r_c = canopy resistance, $s\ m^{-1}$

r_b = canopy bulk lea boundary layer resistance, $s\ m^{-1}$

ρ, c_p = density, specific heat of air; $kg\ m^{-3}$, $J\ kg^{-1}\ K^{-1}$.

γ = psychometric constant, $Pa\ K^{-1}$.

Under normal daytime conditions, $r_c \gg r_b$ so λE_{ct} is proportional to the canopy conductance $g_c = 1/r_c$. Sellers et al. (1992b) showed that for the grass cover found in the FIFE area, g_c could be well-described by

$$g_c = \frac{1}{r_c} = \nabla_F F_0 [f(\delta e) f(T) f(\Psi_L)] \quad (7)$$

g_c	=	canopy conductance, $s\ m^{-1}$
r_c	=	canopy resistance, $m\ s^{-1}$
∇_F	=	$\frac{\partial g_c^*}{\partial F}$, $m^3\ J^{-1}$
g_c^*	=	unstressed (vegetation and light dependent components only) value of g_c , $s\ m^{-1}$
F_0	=	incident flux of PAR, $W\ m^{-2}$

$f(\delta e)$, $f(\Psi_L)$ and $f(T)$ are environmental stress terms which take into account the effects of vapor pressure deficit (δe); leaf water potential $f(\Psi_L)$, which is a function of soil moisture potential and stress; and temperature T .

Equation (7) is similar to the empirical equation set of Jarvis (1976) which was put forward to describe the stomatal conductance of individual leaves as a function of environmental variables. The critical term in (7) is ∇_F which should be a function of FPAR and vegetation physiology only. Figure (5a) shows how

g_c^* , the unstressed or maximum canopy conductance, is supposed to vary as a function of canopy greenness and incident PAR: as the density of green vegetation increases, the area-averaged response of g_c^* to changes in PAR is expected to be more lively, leading to an increase in ∇_F .

The theoretical work of Sellers et al. (1992c) and Sellers (1985, 1987) indicated that ∇_F should be near-linearly related to spectral vegetation indices obtained from satellite data, see Figure (5b). Sellers et al. (1992b) used the FIFE-87 surface flux station data and associated remote sensing data to test the validity of this hypothesis. SiB was adapted to operate in the inverse mode; that is, the model was forced to reproduce latent and sensible heat fluxes that converged on above-canopy flux observations by iterative adjustment of one or two parameters controlling evapotranspiration. This procedure resulted in retrieval of ∇_F values for several surface flux stations for different times of the year which were then compared with spectral reflectances as observed by satellites for the 90m x 90m area (9 Landsat thematic mapper pixels) centered on each flux station. The results are shown in Figure (6) for a pair of flux stations that operated in 1987 and 1989; ∇_F is compared to the simple ratio vegetation index, SR, as given by the ratio of the near-infrared and visible surface reflectances. From Figure (6), we obtain an expression which should be a reasonable descriptor of ∇_F for the C₄ vegetation which dominates the FIFE area.

$$\nabla_F = a \text{ SR} + b; \quad r^2 = 0.717 \quad (8)$$

$$\begin{aligned} a &= 0.495 * 10^{-5}, \text{ m}^3\text{J}^{-1} \\ b &= -1.033 * 10^{-5}, \text{ m}^3\text{J}^{-1} \\ \text{SR} &= \rho_N / \rho_V \\ \rho_N, \rho_V &= \text{near-infrared, visible surface reflectances} \\ &\text{derived from Landsat, SPOT data.} \end{aligned}$$

From (6), (7) and (8), we see that the canopy transpiration rate, λE_{ct} is roughly proportional to g_c , which in turn is near-linearly related to SR. This chain of near-linear relationships suggests that if we were to use area-averaged values of SR to specify $\langle g_c^* \rangle$, there would be little impact on the area-averaged latent heat flux, i.e. $\langle \text{LE} \rangle_M \approx \langle \text{LE} \rangle_I$ in the absence of significant environmental stress. This

is because the transpiration formulation approximately satisfies the condition represented by (4a)

- (iii) Soil Moisture: Soil moisture is a critical regulator of evapotranspiration at the local scale; as soil moisture decreases, vegetation stomates close in response to hormones flowing from the drying root system, see Gollan et al. (1986, 1992) and Mansfield and de Silva (1994). At the same time, the direct evaporation flux from the soil surface rapidly falls off.

The effect of soil moisture stress on transpiration is represented in SiB by $f(\psi_l)$, the leaf water potential stress factor in (7). An expression for $f(\psi_l)$ and the other stress factors, $f(\delta e)$ and $f(T)$ were obtained from curve fits to the data of Verma et al. (1993) such that:

$$f(\delta e) = \frac{1}{1+b\delta e} \quad (9a)$$

$$\begin{aligned} b &= 0.0002 \text{ Pa}^{-1} \\ \delta e &= \text{vapor pressure deficit in the canopy air space, Pa.} \end{aligned}$$

$$f(\psi_l) = \frac{\psi_l - \psi_{c2}}{\psi_{c1} - \psi_{c2}} \quad (9b)$$

$$\begin{aligned} \psi_l &= \text{leaf water potential, Pa} \\ \psi_{c1}, \psi_{c2} &= \text{leaf water potentials at which stomates start to close and are completely closed, respectively, Pa.} \end{aligned}$$

$$f(T) = 1 \quad (9c)$$

The factors $f(\delta e)$, $f(T)$ and $f(\psi_l)$ are allowed to vary between 0, where stress is simulated to completely shut down transpiration, and 1 where no stress limitations are applied to g_c^* .

The leaf water potential, ψ_l , is determined from a catenary equation with the water source defined by the root zone soil moisture potential, a direct function of W_2 , see Sellers et al. (1986). In practice, the leaf water potential is very close to the root zone soil moisture potential

$$\psi_l \rightarrow \psi_2 = \psi_s W_2^{-b} \quad (10)$$

ψ_2	=	root zone (second soil layer)
		soil moisture potential, Pa
ψ_s	=	saturated value of ψ_2 , Pa
W_2	=	root zone soil wetness fraction
	=	1 when saturated
b	=	sorptivity parameter

The soil hydraulic parameters ψ_s , b , soil porosity and soil hydraulic conductivity are all functions of soil type. With regard to $f(T)$, the work of Verma et al (1993) and Polley et al. (1992) indicated that leaf photosynthesis was only weakly dependent on leaf temperature in the C4 species found in the FIFE area and so $f(T)$ was set to unity.

Soil evaporation can make a significant contribution to the total evapotranspiration flux, see Villalobos and Fereres (1990) and Sellers et al. (1992b). In Sellers et al. (1992b) a series of inverse-mode runs with SiB were used

to estimate soil evaporation rates and soil surface resistances, r_{surf} , from the time-series of flux station data: this procedure resulted in the formulation of an empirical expression relating r_{surf} to the wetness of the uppermost soil layer, W_1

$$r_{\text{surf}} = \exp(8.206 - 4.255W_1) \quad (11)$$

$$\begin{aligned} r_{\text{surf}} &= \text{soil surface resistance, } \text{sm}^{-1} \\ &\geq 150 \text{ sm}^{-1} \\ W_1 &= \text{soil wetness of the (0-5cm) upper soil layer.} \end{aligned}$$

Figure (7) shows the variation of $f(\Psi_L)$ and r_{surf} with soil wetness (W_2 and W_1 , respectively). Clearly, these functions are very non-linear for dryer soil conditions. It is also widely reported that soil moisture variability can be significant, particularly for wet soil moisture conditions, see for example Charpentier and Groffman (1992), tables 3 and 4. We might therefore expect that since neither (4a) or (4b) are likely to be satisfied by the soil moisture-surface flux relationships, then area-averaging soil moisture over an intermediate scale domain could have severe impacts on the calculated fluxes, i.e. $\langle Q \rangle_I \neq \langle Q \rangle_M$.

3. The FIFE Data and Model Validation at the Local Scale

3.1 The FIFE-89 Data Set

The FIFE-89 'testbed' data set was selected for the study of Sellers et al. (in prep.) which focused on the impacts of spatial variability on area-averaged fluxes. Much of the work reported in Sellers et al. (in prep.) is summarized here.

The data collection effort in FIFE-89 was concentrated primarily within a 2 x 15 km strip oriented north-south in the middle of the FIFE area. This so-called 'testbed' area contained most of the surface flux stations and was also made the first priority for airborne data collection.

There was considerable spatial variability in the fields of surface soil moisture and green vegetation cover in FIFE-89 due to the after-effects of convective storms that covered the southeastern corner of the area on days 204 and 212. In addition, the area was subjected to a long dry period from day 216 through day 224 during which a series of airborne remote sensing missions collected microwave data to monitor the drying-out of the surface soil layer. Airborne and surface flux measurements tracked the decrease in the surface latent heat fluxes over the same period. All in all, the FIFE-89 data set gives us a good example of a landscape with well-developed spatial heterogeneity in topography, vegetation condition and soil moisture with some significant changes in soil moisture over time as the site dried out. It should be remembered, however, that the vegetation type is an almost uniform grassland within the domain.

The critical data collected in FIFE-89 are depicted schematically in figure 8. They include:

- (a) Surface boundary conditions and model initialization data
- (i) Atmospherically-corrected SPOT and Landsat TM surface spectral reflectances at 30 x 30m, combined into SR fields.

variability in these forcings during the dry period, day 216 through 224, and so spatial means of the AMS variables were used to provide uniform upper boundary conditions for the model runs.

(c) Model validation data

(vi) Surface fluxes: The surface flux stations shown in Figure 8 provided near-continuous time-series estimates of net radiation, latent heat, sensible heat, momentum and in some cases total CO₂ flux. The instruments and techniques are described in Kanemasu et al (1992).

(vii) Airborne fluxes: The National Research Council of Canada contributed the Twin Otter flux aircraft to both FIFE-87 and FIFE-89. The aircraft is equipped to measure the turbulent fluxes of sensible and latent heat, CO₂ and momentum, see Desjardins et al. (1992a), and MacPherson et al. (1992).

3.2 Model Initialization

The Landsat 30 x 30m data grid as defined for the FIFE area by FIS was used as the basis for the full-resolution studies within the FIFE-89 testbed area. The digital elevation model (DEM), soil moisture and other data sets were all gridded to this 68 x 501 pixel domain. The full-resolution calculations of energy and water balances discussed in the next section were performed for each of these pixels.

- (ii) Spatially-continuous airborne microwave data collected by the NASA C-130 aircraft equipped with the push broom microwave radiometer (PBMR). The PBMR data, described in detail in Wang et al (1992), were used to infer surface (0-5cm) soil moisture fields using a regression equation based on a large number of gravimetric samples acquired at the same time as the PBMR data. The PBMR data were mapped by Wang et al. (pers. comm.) to a 15 x 15m grid.
- (iii) Soil moisture profiles. In addition to the daily near-surface soil gravimetric samples taken around the flux station sites, the soil moisture profile from 10 cm to the bedrock (usually at 50 to 150 cm depth) was monitored using neutron probes.
- (iv) Digital Evaluation Model (DEM). A 25m resolution DEM was prepared from U.S. Geological Survey data by the FIFE Information System (FIS). The vertical resolution of these data appear to be good to better than 5m, see Davis et al (1992).

(b) Model forcing data

- (v) The network of automatic meteorological stations (AMS) and instruments mounted on the surface flux stations provided continuous measurements of the near-surface meteorological variables; temperature, mixing ratio, wind speed (T , q , u); downwelling shortwave, longwave and net radiation fluxes ($S\downarrow$, $L\downarrow$, R_n); and precipitation (P) over the area. Some comparisons showed that except for precipitation there was very little spatial

The ∇_F , FPAR and leaf area index (LAI) values for each pixel were specified from SR fields as derived from TM or SPOT data using equation (8), and methods discussed in Sellers et al (1992a, b). The LAI, canopy greenness fraction and soil background reflectance values are used by the SiB radiation submodel to calculate surface albedo and the absorption of radiation by the canopy and ground. Likewise, these LAI data enter into the turbulent transfer submodel to calculate roughness lengths (z_0) and other aerodynamic parameters. Soil physical properties (porosity, Θ_s ; saturated hydraulic conductivity, K_s ; saturation moisture potential, ψ_s ; sorption parameter, B) are specified from data collected by the FIFE Staff Science soil survey work and the figures of Clapp and Hornberger (1978), see Sellers et al (1992b).

The prognostic variables of SiB are the canopy, ground surface and deep soil temperatures (T_c , T_g and T_d ; respectively), the canopy and soil interception stores (M_c , M_g) and the soil moisture contents for the surface (0-5cm) layer, W_1 ; the root zone, W_2 ; and the deep soil moisture store, W_3 .

T_c and T_g are initialized from the air temperature and T_d is set equal to the 50cm soil temperature as measured by the AMS network; mean values are used for the entire testbed area. M_c and M_g are both set to zero for the first timestep.

The surface layer (0-5 cm) soil wetness, W_1 , was initialized using the PBMIR data of Wang et al (1992). The soil moisture contents for the root zone, W_2 , and the deep soil layer, W_3 , are more difficult to specify directly from observations; all we have are the neutron probe readings for the AMS and flux station sites. Using these and the other available FIFE-89 data, a study was made into which of the following variables are the best predictors of W_2 and W_3 : SR , W_1 (from the

PBMR data), slope and elevation. The best fits were provided by regression equations operating on local slope angle, simple ratio and W_1 ; see Sellers et al. (in prep.).

The deepest neutron probe reading taken at AMS and flux station sites generally corresponded to the bedrock depth. These readings were compared to the local slope and elevation data to specify the spatial variation of the soil profile depth, see Sellers et al. (in prep.): This depth varied from 0.25 to 2.0m in the testbed area.

3.3 Model Operation

Generally, model runs were started at some time close to the acquisition of the PBMR data used to initialize W_1 . This is because all of the other initialization data are more-or-less continuously available (temperatures), slowly changing in time (∇_F , FPAR and LAI) or constant (soil properties, topography). For the full resolution calculations, the energy and water balances were calculated at hourly-time resolution for each of the $(68 \times 501 = 34068)$ 30×30 m pixels within the testbed domain.

With the incorporation of the modified radiation code, see (5), SiB can be used to compare full resolution (area-integrated) with block (area-averaged) calculations of the surface heat fluxes.

Area-integrated values of prognostic or diagnostic variables calculated by these full-resolution runs are provided by simply taking the mean of the appropriate quantities over all the pixels in the testbed domain, see the left-hand side of

figure 9. Area-averaged calculations are performed by performing a single model run using area-averaged initial conditions and prognostic variables, see the right-hand side of figure 9. The initial conditions for the area-averaged runs are provided by averaging the initial fields used by the full-resolution runs. A mean of the absolute values of the slopes of all the pixels is used to calculate the baseflow from the soil recharge layer (W_3). For the radiation calculations, the area-averaged slope and aspect over the domain work out to be almost negligible.

3.4 Model validation using surface and airborne flux data

The $90 \times 90\text{m}$, 3×3 pixel cluster centered on each flux site is assumed to be representative of the surface flux 'footprint' characterized by that flux station, see Schuepp et al. (1990). Figure 10a shows model time-series of fluxes produced by the SiB model for site 916, day 216 using the initial conditions and parameter sets specified in the previous section; the model does a reasonable job of reproducing the observed fluxes. Figure 10b shows the results of model calculations for a number of flux stations for a day with relatively wet soil moisture conditions, day 216, and days with dryer soil moisture conditions, days 222, 223 and 224. The evaporative fractions shown in Figure 10b are the means of the calculated and observed fluxes for the periods when the Twin Otter flux aircraft was flying over the area. The results from the same full-resolution runs were area-integrated to provide estimates for 2×2 km blocks for the same periods, for comparison with overlapping 2×4 km blocks of footprint-adjusted surface flux estimates produced by the Twin Otter, see Figure 10c and Desjardins et al. (1992a). The flux data of Desjardins et al. (1992a) are not directly comparable to the calculations as the domains are different (2×4 km versus 2×2 km) and it is

known that estimates of the surface fluxes provided by the flux aircraft are consistently low, see Desjardins et al. (1992b) and Sellers et al (1992b). However it appears that ratios of these estimates, eg the evaporative fraction, $EF = \lambda E / (\lambda E + H)$, are more consistent with equivalent surface observations, see Figure 10c and Sellers et al. (1992b).

The results shown in Figure 10 and other work reported in Sellers et al (1992b; in prep.) indicate that all aspects of the model perform credibly at the local (90 x 90m) scale when using initial conditions provided by meteorological observations for the prognostic variables, and remotely sensed measurements for the vegetation condition (∇F , albedo, etc.) and surface soil wetness (W_1). The model also appears to reproduce the responses of the energy balance to changes in soil moisture at this spatial scale, as can be seen by the range of EF values in Figure 10.

4.0 Investigating the Effects of Inhomogeneity on Intermediate (2 x 15 km) Scale Fluxes

4.1 Exploratory tests

Three kinds of surface heterogeneity are addressed in this study: topography (slope and aspect); vegetation condition (∇F and LAI); and soil wetness (principally variations in W_1). Table 1 and figure 11 show the results of exploratory tests performed using the FIFE-89 testbed data set. For days 216 and 224, a series of 24-hour simulations were performed starting with a full resolution calculation in which all these fields were fully resolved, followed by a succession of runs in which one or more of the initial fields was assigned a mean value over the whole domain. The final rows of Tables 1a and 1b show the results

of the area-average 'block' runs where a single simulation is done for the entire testbed area using area-averaged initial conditions and prognostic variables. This area-average calculation is identical to the area-integral of pixel-scale calculations carried out over the testbed domain where each pixel is given the same mean initial and boundary conditions.

Figure 11 shows the diurnal course of the heat fluxes for the testbed domain as calculated using fully-resolved and area-averaged fields of soil moisture, topography and SR. It is surprising how closely the results from the two methods agree, see also figure 14 of Sellers et al. (1992b).

The results in Table 1 show what we would expect from the analysis in section 2.0 of this paper. The functions governing the response of the energy balance to changes in vegetation condition and topography are more-or-less linear in these variables, so the impact of area-averaging those variables is weak. This result was anticipated for the vegetation component by Sellers et al. (1992b) and for the effects of topography on the radiation balance by Dubayah et al. (1990).

However, the dependances of the soil surface resistance, r_{surf} , on W_1 , and the vegetation moisture stress function, $f(\Psi_L)$, on W_2 are non-linear, see Figure 7. As a result, the area-averaging of soil wetness fields leads to a relatively large divergence between the area-integrated and area-averaged surface flux estimates in Table 1. A comparison of lines 1 (all three fields fully resolved) and lines 5 (only the soil moisture field fully resolved) in Tables 1a and 1b reveals that soil wetness variability accounts for most of the calculated variability in the latent and sensible heat flux fields ($\sigma_{\lambda E}$ and σ_H) and, more significantly, for the greatest

mismatch between the area-integrated (line 1) and area-averaged (lines 5 and 8) estimates of heat fluxes.

Sellers et al. (in prep.) developed an index which quantified the differences between $\langle Q \rangle_I$ and $\langle Q \rangle_M$ relative to standard deviations in the surface property fields. This index shows clearly that soil moisture is the worst-behaved of the three variables with respect to area-averaging. Given these findings, it may be concluded that simple area-averaging procedures can be used on the fields of SR (provided we are dealing with one type of vegetation in the domain) and topography without greatly impacting the calculated larger-scale fields of sensible and latent heat flux. In the next section, we therefore concentrate on the influence of spatial variability in the soil wetness field on the mean surface heat fluxes.

4.2 Sensitivity tests: the impact of soil wetness variability on the surface heat fluxes

4.2.1 Case I: Test with constant spatial variability in the soil wetness field

When we run a simulation forward from a set of initial conditions using observed meteorological forcings, it is often difficult to separate a clear signal of the impact of changes in soil moisture variability on the surface fluxes from other effects; e.g., changing weather conditions, changes in the mean soil moisture condition, etc. To isolate the effects of soil moisture variability from other effects, the day 216 initial conditions and meteorological forcings were used as a baseline. To start with, reference runs were done with fully-resolved and area-averaged fields of topography (top), soil wetness (W) and SR for day 216; the

results of these runs are shown in lines 1 and 8 of Table 1a and in Figure 12 where the area-averaged evaporative fraction (EF) is plotted against the average soil wetnesses W_1 and W_2 for the testbed domain. Subsequent runs, which used both fully-resolved (top, W, SR) and area-averaged ($\langle \text{top} \rangle, \langle W \rangle, \langle SR \rangle$) initial conditions, were performed using the same forcings and initial conditions with the sole exception that the entire soil wetness field, i.e., all pixel-level values of W_1 , W_2 and W_3 , were raised or decreased by a constant fraction in each run. For example, if we represent the wetness of a pixel on day 216 by W^{216} , subsequent runs in the series would start with values for that pixel of $W^{216} - 0.05$, $W^{216} - 0.10$, $W^{216} - 0.15$ and also with $W^{216} + 0.05$, $W^{216} + 0.10$, etc. This procedure has the effect of preserving the soil moisture variability and forcing conditions for the testbed area while adjusting the mean soil wetness. (Individual pixel values of W were not permitted to drop below 0.05 or exceed 1.0 during this test; the values of $\langle W \rangle$ plotted in Figure 12 are the means taken over the W field following these bounding adjustments so that at low and high values of $\langle W \rangle$, successive increments and decrements of $\langle W \rangle$ may be slightly less than 0.05). Figure 12 shows that for the day 216 pattern of soil wetness, the fully-resolved and area-averaged calculations of EF are close to each other at high $\langle W \rangle$ but diverge as $\langle W \rangle$ decreases. Figure 7 showed why this is: the soil moisture stress function, $f(\Psi_t)$, and the soil surface resistance function, r_{surf} , exert little control on evapotranspiration when soil moisture is freely available but have large non-linear influences as W decreases.

The results in Figure 12 (Case I) indicate that these non-linear effects become a problem when W drops below about 0.5. However, it should be remembered that the spatial variability of W in Case I was artificially preserved as W decreased in this test. Are these patterns conserved in nature?

4.2.2 Case II: Changes in soil moisture variability during a real drydown

Figure 13 (Case II) shows the result of running fully-resolved and area-averaged versions of SiB forward from day 216 through day 224 using the observed time-series of meteorological forcings and starting with the day 216 initial conditions. Figures 13a and 13b show time-series of calculated available energy ($R_n - G$) and evaporative fractions (EF) over the period; Figure 13c and 13d show the changes in the surface and root zone soil moisture. If anything, the variability in EF is conserved over time. Also shown in the figures are observations of some of these quantities obtained by averaging available surface and airborne heat flux measurements. While the figure indicates that the model is performing credibly, figure 13c is particularly interesting: the spatial variability in the surface layer wetness, σ_{W_1} , is seen to decrease as W_1 decreases, see also Figure 13e which shows the same data replotted for clarity. The likely explanation for this trend is that wetter areas tend to dry out and/or drain much quicker than dryer ones which has the effect of continually narrowing the range of soil moisture contents within the domain as the drydown progresses. The results in Figure 13 can be summarized as follows:

- (i) The model appears to simulate the changes in soil wetness and heat fluxes reasonably well over the FIFE-89 drydown period (day 216-224).
- (ii) As the domain dries out, the spatial variability in evaporative fraction seems to be more or less invariant.

- (iii) As the domain dries out, the spatial variability of the surface and root zone soil wetness decreases.

4.2.3 Case III: Changes in the surface energy balance and soil wetness fields during an idealized drydown

The results from Case II are contaminated by day-to-day changes in the meteorological forcings; for example, the observed and calculated evaporative fractions in Figure 13b rise and fall in response to the weather, partially obscuring the overall downward trend in EF that we would expect to see as the soil moisture decreases.

Figure 14 (Case III) compares the results of fully-resolved and area-averaged SiB runs starting from day 216 initial conditions but forced by repeated cycles of the day 216 meteorological conditions for every day thereafter. Figures 14a, b and c correspond to Figures 13a, b and c. It can be seen that the time-series of energy balance components, EF and the average soil wetness calculated by the two methods track each other very closely over the course of the drydown. This is not what we might expect from Case I (Figure 12) which indicated that the two methods would produce increasingly divergent surface flux estimates as W decreases. However, it is clear that the reduction in σW as W decreases counteracts this divergence with the net result that the two methods produce very similar time-series of area-averaged surface fluxes and soil wetnesses.

Figure 15a shows the variation in σW_1 as a function of W_1 for all three cases. The uppermost line represents Case I where σW_1 was effectively conserved (see Figure 12). The solid line and data points represent the modeled (full resolution) and observed (from the PBMR data) values of σW_1 for Case II, the real drydown

case for days 216 through 224 shown in Figure 13. The dashed line shows the results of Case III, the idealized (cyclically repeated day 216 meteorological forcings) drydown. Clearly, Case I is unrealistic. These Case II and Case III calculations are supported by the available data, also shown in Figure 15b.

From these results, we conclude that:

- (i) As the soil moisture field dries out, its spatial variability is reduced. This has the effect of reducing variabilities in the surface heat flux fields below levels they would maintain if σ_w were conserved.
- (ii) Because of the above effects, the area-averaged and area-integrated estimates of the surface heat fluxes and soil wetness track each other closely during a drydown.

5.0 Summary

Soil-Vegetation-Atmosphere Transfer (SVAT) models are usually developed from formulations which are conceptualized and tested at local scales, a few meters to a few tens of meters. It is the normal practice to apply these models directly as land surface parameterizations (LSPs) in atmospheric general circulation models (GCMs), a procedure which implicitly assumes that the dependence of the land-atmosphere fluxes on critical land surface parameters, principally topography, vegetation condition and soil moisture, are scale-invariant. Analyses in the second section of this paper demonstrate that for this assumption of scale-invariance to hold true, either the models linking fluxes to surface conditions must be linear or nearly so, or the spatial variability in the

critical surface initial/boundary conditions must be small. The analyses suggests that the negative impacts of area-averaging surface fields would be relatively small for topography and vegetation condition but could present some problems when dealing with soil moisture. This analysis provided the motivation for a numerical study of the issue using a modified version of the Simple Biosphere model (SiB) of Sellers et al. (1986) and the FIFE-89 data set.

The third section evaluated the performance of SiB using initial/boundary conditions (vegetation condition, represented by the conductance parameter, ∇_F , from SR; soil moisture from airborne microwave and in situ data); meteorological forcings (from the automatic meteorological station network); and validation data (surface and airborne flux measurements) gathered for the 2 x 15 km testbed area located in the center of the FIFE area. The model heat flux simulations agreed convincingly with available observations over a range of soil moisture conditions.

The fourth section addresses the central question of the paper: how does spatial variability in topography, vegetation conditions (∇_F) and soil wetness affect the area-averaged fluxes? The approach for investigating this issue was to perform 'fully-resolved' calculations for every 30 x 30m pixel within the testbed area, i.e. for a total of 68*501 pixels. This 30m grid is fine enough to resolve landscape-scale variability in the area and also values for the fields can be assigned at this resolution from the available observations. These full-resolution runs were followed by similar runs in which one or more of the spatially-varying fields was replaced by a domain-average value. It was found that the use of area-averaged fields of topography and the vegetation index, SR, had relatively small impacts on the mean fluxes. This is mainly because the relations linking topography and

SR to the surface energy and heat fluxes are near-linear and so area-averaging and area-integration operations are mathematically almost equivalent. Some of these runs (Case I) indicated that if a large spatial variability in the soil wetness field could be maintained as the area dried out, the area-averaged and full-resolution (area-integrated) flux estimates would be significantly different. This is because the relationships between soil moisture content and soil and canopy evapotranspiration become increasingly non-linear as the soil dries. Further simulations tracked the dynamics of the soil wetness field during the FIFE-89 drydown (days 216 through 224) for which comprehensive data sets were available. Both the simulations and observations indicated that as the soil wetness decreased, its spatial variability also decreased with the result that the area-averaged and area-integrated calculations of the surface fluxes tracked each other closely as the drydown progressed. In short, when the soil is wet, the soil wetness field has high spatial variability but the evapotranspiration rate is insensitive to these variations for these conditions. By contrast, when the area-averaged soil wetness is low, its spatial variability is also low and so even though the soil wetness-evapotranspiration functions are highly non-linear under these conditions, their impact on the area-averaged fields is small. These findings are summarized in figure 16. In conclusion:

- (i) The relationships describing the effects of moderate topography on the surface radiation budget are near-linear and so are largely scale-invariant. The relationships linking the simple ratio vegetation index (SR), the canopy conductance parameter (V_F) and the canopy transpiration flux are also near-linear and similarly scale-invariant to first order. Because of this, simple area-averaging operations can

be applied to these fields with relatively little impact on the calculated fields of the surface heat fluxes.

- (ii) The relationships linking surface and root zone soil wetness to the soil surface and canopy transpiration rates become increasingly non-linear as the soil dries out. However, simulation results and observations both showed that soil wetness variability decreases significantly as an area dries out, which partially cancels out the effects of these non-linear functions. As a result the time-series of heat fluxes produced by fully-resolved and area-averaged soil wetness fields track each other closely.

The combination of the effects described in (i) and (ii) above indicate that we can use area-averaged values of topography, vegetation condition and soil wetness to describe fluxes on the scale of a few kilometers over a period of a few days. They also indicate that for practical purposes, the biophysical evapotranspiration models in use for calculating these fluxes are relatively robust and scale-invariant with respect to these variables. It should be noted that these results apply only to the calculation of surface flux fields; we suspect that such simple area-averaging techniques will not work well in describing runoff rates, particularly under very wet soil moisture conditions which were not addressed here. How much confidence do we have that these findings translate to the larger scales associated with atmospheric models, i.e. 50 - 500 km? The results of this paper and those in Sellers et al (1992b) indicate that simple area-averages of (moderate) topography and vegetation parameters are probably sufficient over a wide range of spatial scales provided that the vegetation type does not change radically within the domain; the work of Noilhan et al (1991) supports this view. However, the issue

is not so straightforward with respect to soil moisture. To address it properly, it would be necessary to characterize the dynamics of the spatial variability of soil moisture and atmospheric conditions at these larger scales. If we see the same trends of decreasing spatial variability in soil wetness with decreasing area-averaged wetness, then the entire biophysical modeling approach can be treated as effectively scale-invariant over the spatial scale range of a few meters to a few tens or hundreds of kilometers. If this turns out not to be the case, then scale-dependent area-averaged forms of the equations relating soil moisture to evaporation will have to be developed which take these effects into account. However, it must be said that the results of this study suggest that the impact of using area-averaged values of soil moisture, vegetation cover and topography on the calculated fluxes of sensible and latent heat are much less than were originally feared by the authors. This finding in turn suggests that inaccuracies associated with the assumption of biophysical model scale-invariance are likely to be small.

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Table 1: Results of exploratory tests described in section 4.1. Initial/boundary conditions and forcings for (a) day 216 and (b) day 224 were used to run SiB for the testbed domain. The results of model runs with fully resolved fields of topography (top), soil moisture (W) and simple ratio (SR) are shown in lines 1 of Tables 1a and 1b; line 8 shows equivalent results when all three fields are represented by area-averages which degenerates to a single energy balance calculation for the entire domain. Lines 2 through 7 show the results of using area-averages of one or two of the three fields.

Figure Captions

Figure 1

Schematic diagram of the Simple Biosphere model (SiB) as modified for use in Sellers et al. (1992a) and this study. This version differs from the original one published in Sellers et al. (1986) in that the ground vegetation cover has been removed and the FIFE site grass canopy has been "promoted" to be the upper story vegetation layer. The fluxes of sensible and latent heat flux from the surface to the atmosphere are depicted on the right-hand and left-hand sides of the figure, respectively.

Figure 2

Left hand side: A model relating flux Q at a point to a surface variable x_i (e.g. soil moisture) can be numerically integrated over a domain to provide an accurate estimate of the domain-scale flux, $\langle Q \rangle_I$. In practice, this is done by dividing the domain into many small-scale pixels where the length scale is known to be compatible with the model, $Q = f(x_i)$, and summing the results of the individual pixel scale calculations.

Right hand side: The same model can be applied to area-averaged surface parameter values, $\langle x_i \rangle$ to yield an estimate of the area-averaged flux $\langle Q \rangle_M$. However, if $f(x_i)$ is non-linear and the spatial variability of x_i is significant, then $\langle Q \rangle_M$ will be different from the 'true' value as defined by $\langle Q \rangle_I$.

Figure 3

Effect of using area-averaged values of surface properties x_i to calculate surface fluxes, $Q = f(x_i)$.

- (a) For a linear function $f(x_i)$, there is no difference between a full-resolved integration of $Q=f(x_i)$ and a calculation based on $\langle Q \rangle = f(\langle x_i \rangle)$; the geometry in this figure explains why this is so. The area-integrated calculation, $\langle Q \rangle_I$, corresponding to the left hand side of Figure 2, is represented by a solid circle '•'; the area-averaged calculation, $\langle Q \rangle_M$, corresponding to the right-hand side of Figure 2, is represented by an open circle 'o'.
- (b) For a non-linear function $f(x_i)$, there can be a significant difference between $\langle Q \rangle_I$ and $\langle Q \rangle_M$ where the spatial variability in x_i is large (upper case). When the spatial variability in x_i is low, $\langle Q \rangle_M \rightarrow \langle Q \rangle_I$.

Figure 4

Interception of (a) direct beam radiation and (b) diffuse radiation as treated in this modified version of SiB. The factor β_b in equation (5a) in text represents the ratio of radiation intercepted by a sloping pixel compared to a horizontal one. In (a), the solar beam has the same azimuth as the slope and β_b is equivalent to the area shadowed by the pixel $(1+x)$. In (b) each portion of the pixel views a fractional area of sky above the horizon of $(\pi-V)/\pi$ and an area of landscape below the horizon of fraction area V/π .

Figure 5

Schematic figures illustrating the hypothesized relationships between the unstressed canopy conductance, g_c^* , and the simple ratio (SR) vegetation index. (a) Changes in unstressed canopy conductance, g_c^* , for a range of canopy states and incident photosynthetically active radiation (PAR) fluxes, F_0 . Note that all the canopy conductances saturate at the same level of F_0 , F_0^{sat} , that is, $V_{\text{max}0}$ is invariant in this example. Derivatives of g_c^* with respect to F_0 , ($F_0 < F_0^{\text{sat}}$) are shown. (b) Derivatives of g_c^* with respect to F_0 , $V_F \equiv \partial g_c^* / F_0$, are plotted against SR vegetation index. The letters on the line correspond to the cases shown in Figure 5a. These figures illustrate the principles discussed in the text; they contain no real data.

Figure 6

Retrieved values of V_F for one FIFE-89 and two FIFE-87 flux stations plotted against remotely sensed variables derived from Landsat TM (solid lines) and SPOT (dashed lines) data. Data for the FIFE-87 stations 16 and 18, and for the FIFE-89 916 station are shown (sites 16 and 916 are exactly equivalent, sites 16 and 18 were collocated). The V_F values are plotted against Landsat TM and SPOT reflectance-based SR values; error bars represent the uncertainties due to the soil moisture formulation, and the overall model goodness-of-fit.

Figure 7

Dependence of the soil surface resistance, r_{surf} , on W_1 (equation 11) and the vegetation moisture stress function, $f(\Psi_t)$, on W_2 (equation 9b).

Note that: (i) r_{surf} is kept constant for values of $W > 0.75$. In practice, such high levels of surface soil moisture cannot be maintained for very long (ii) $f(\Psi_t)$ is plotted as a direct function of W_2 for the purpose of illustration; i.e. the plant and soil resistance terms are taken as negligible here, so $\Psi_t = \Psi_2$.

Figure 8

Schematic showing the FIFE-89 testbed area data sets used in this study. The figure also shows the location of the testbed area in the FIFE area.

A. Initial/boundary conditions

- (i) Vegetation condition: SPOT and TM data were used to calculate surface reflectances (ρ_N , ρ_V) and simple ratio vegetation index (SR) fields at 30 x 30 m resolution.
- (ii) Surface soil moisture: The airborne pushbroom microwave radiometer (PBMW) mounted on the NASA C-130 mapped surface (0-5 cm) soil moisture at 50 x 50 m resolution.
- (iii) Root zone, deep soil moisture: Neutron probe and gravimetric data were used to develop regression equations to specify these soil moisture fields.
- (iv) Topography: Digital Elevation Model (DEM) from USGS data, aggregated to 30 x 30 m resolution.

B. Forcing variables

- (v) Near-surface meteorology and downwelling radiation fluxes: Automatic Meteorological Stations (AMS) measured near-surface meteorological variables and downwelling shortwave and longwave radiation fluxes.

C. Validation data

- (vi) Surface fluxes: surface flux stations measure fluxes at half-hourly time resolution.
- (vii) Airborne flux measurements: The Canadian Twin Otter aircraft flies at 15 m altitude to map surface fluxes at coarse (2 x 4 km) resolution.

Figure 9

Schematic comparing full-resolution (68 x 501 pixels) and area-averaged 'block' calculations for the FIFE-89 testbed area. In this example, all the fields represented in detail in the full-resolution calculation are area-averaged to provide a set of mean initial/boundary conditions for the area-averaged run, which can then be treated as one big pixel. Any intermediate calculation,

where at least one field is represented at full resolution, must be carried out at full resolution.

The full resolution flux fields may be area-integrated or aggregated for direct comparison with the area-averaged results.

Figure 10
Comparisons between surfaces fluxes calculated by SiB in this study and as measured in the field in FIFE-89.

- (a) Comparison of observed and calculated diurnal courses of surface heat and energy fluxes for site 916 (ECV-4439) for day 216. Calculated heat fluxes for a 90 x 90 m area centered on the flux station site (3 x 3 Landsat pixels) are shown.
- (b) Comparison of observed and calculated evaporative fractions for a number of surface flux stations in FIFE-89; the numbers on the figure represent individual sites. The points above the dashed line cutting the 1:1 line are for the period 1345 - 1545 local time, day 216 (relatively wet soil conditions) and the points below the dashed line are for the periods 1045 - 1215, day 222; 1145 - 1345, day 223; and 1145 - 1345 local time, day 224 (relatively dry soil moisture conditions).
- (c) Comparison of observed and calculated evaporative fractions for a series of blocks aligned along the center of the testbed area, see figure 8. The observed values were produced for 2 x 4 km blocks from airborne eddy correlation data. The calculated values represent 2 x 2 km blocks centered on the airborne data blocks; they were 'averaged-up' from the 30 x 30 m calculations for the same periods and days specified in (b) above. The dashed line delineates between day 216 and day 224 results.

Figure 11
Comparison of energy balance calculations conducted for the FIFE-89 testbed area using the full-resolution (solid line with error bars) and area-averaged or 'block' (heavy dashed line) methods. The full-resolution results represent the mean and standard deviations of the 68 x 501 pixel calculations within the domain; the area-averaged results are based on a single 'big-pixel' calculation for the entire testbed area using mean initial/boundary conditions and prognostic variables.
Day 216: (a) latent heat flux (b) sensible heat flux.
Day 224: (c) latent heat flux (d) sensible heat flux.

Figure 12
Results of the Case I simulation. Plot of the mean (spatial average over the FIFE-89 testbed area; time-average of fluxes over two hours centered on midday) evaporative fraction (EF) as a function

of mean initial soil wetness. Solid lines, with error bars showing standard deviations of EF, represents the results of the full-resolution runs (top, W, SR); dashed lines represent the results of runs where area-averaged initial conditions ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle \text{SR} \rangle$) were used. These results were obtained by taking the initial day 216 soil wetness field as a reference (marked with an arrow) and incrementing/decrementing the wetness of each pixel by successive multiples of ± 0.05 . (a) EF versus W_1 ; (b) EF versus W_2 .

Figure 13

Results of the Case II simulations. Time series of FIFE-89 testbed simulations for the period day 216 through day 224. The solid lines, with light dashed lines showing standard deviations, represent the results of full resolution runs (top, W, SR); the heavy dashed lines show the results of the area-averaged ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle \text{SR} \rangle$) runs. In all cases, mean values for the entire testbed area are shown, time-averaged for two hours centered on local midday. (a) Rn-G; (b) Evaporative fraction (EF); (c) W_1 ; (d) W_2 ; (e) standard deviation of W_1 , σ_{W_1} , plotted against the area-average of W_1 , $\langle W_1 \rangle$. Observations are also shown; heavy circles and solid error bars are for the surface flux stations in the testbed area; open circles and dashed error bars are for the airborne fluxes; crosses with numbers are estimates of σ_{W_1} calculated from surface soil moisture observations derived from the PBM data for each day.

Figure 14

Results of the Case III simulations. Time-series of FIFE-89 testbed simulations starting with day 216 initial conditions and forced thereafter with repeated day 216 meteorological and radiation data. The solid lines, with error bars showing standard deviations, represent the results of the full-resolution runs (top, W, SR); the dashed lines show the results of the area-averaged ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle \text{SR} \rangle$) runs. In all cases, mean values for the entire testbed area are shown, time-averaged for two hours centered on local midday. (a) Rn-G; (b) Evaporative fraction (EF); (c) W_1 ; (d) W_2 , (e) Standard deviation of W_1 , σ_{W_1} , plotted against $\langle W_1 \rangle$ from the full-resolution run and, (f) hourly depiction of W_1 (top, W, SR) for first seven days, again from the full-resolution run: note diurnal variation in W_1 due to daytime evaporation and night time recharge from below.

Figure 15

Variation in σ_{W_1} as a function of $\langle W_1 \rangle$ for Case I (light dashed line); Case II (heavy dashed line); and Case III (solid line). Estimates of σ_{W_1} obtained from analyzing the PBM data are also

shown labeled by day number; these should be compared with the Case II trajectory.

Figure 16

Conceptual diagrams showing the relationships between the value of a surface initial/boundary condition and the evaporative fraction (EF). (a) Topography, represented by the absolute value of the slope; (b) Vegetation condition, represented by variations in the canopy conductance parameter, ∇_F ; and (c) soil wetness, W . In all three figures, the estimate of EF obtained from an area-averaged value of the initial/boundary condition is represented by a solid circle; the area-integrated estimate of EF, which is partly a function of the standard deviation of the initial/boundary condition is represented by an open circle. The two estimates converge when the functions are linear or the standard deviations are small. Note that since the relationships with slope and ∇_F are more-or-less linear, the use of area-averaged values of these variables over an intermediate-scale (2-10km)² domain has little impact on the area-averaged fluxes and EF. The relationship between W and EF is near-linear at high and very low values, so area-averaging W has little impact under these conditions. During a drydown, intermediate values of W are produced in the near-linear portion of the W -EF relationship; however, the dynamics of the drying process act to reduce the spatial variability of W , σ_W , so once again the impact of averaging W is small.

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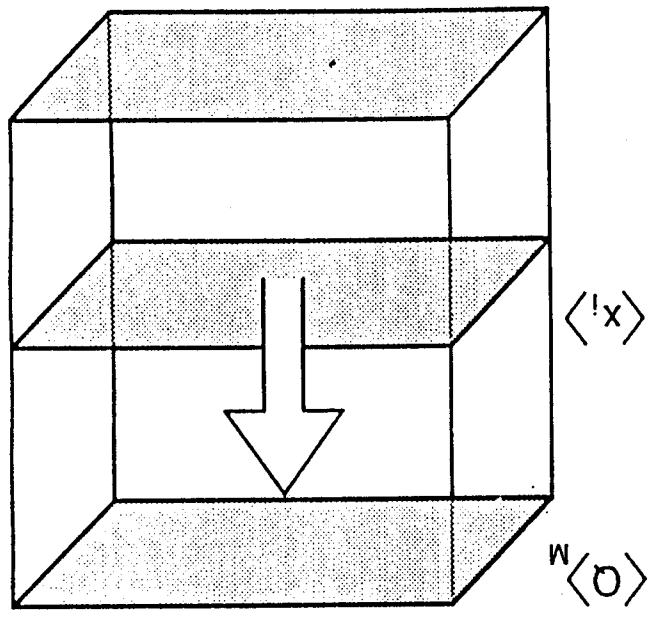
Table 1

(a)

i	treatment	R _n -G		λE		H		G		EF	
		mean	σ	mean	σ	mean	σ	mean	σ	mean	σ
1	top SR W	404.6	32.4	272.4	52.8	132.5	54.0	40.8	3.3	0.653	0.092
2	<top> SR W	407.4	13.5	273.6	51.7	133.8	50.8	40.8	2.3	0.652	0.091
3	top <SR> W	404.0	32.0	270.4	47.6	133.9	45.2	40.8	2.6	0.651	0.069
4	top SR <W>	404.4	32.0	270.8	31.5	133.8	39.2	41.0	3.1	0.647	0.069
5	<top> <SR> W	407.1	13.1	271.7	46.1	135.4	41.9	40.9	2.0	0.650	0.069
6	<top> SR <W>	407.7	12.7	272.1	29.9	135.1	34.5	41.0	2.6	0.646	0.069
7	top <SR> <W>	404.0	31.7	269.7	12.6	134.5	19.3	41.0	2.0	0.647	0.110
8	<top> <SR> <W>	407.6	0.0	271.1	0.0	136.5	0.0	41.2	0.0	0.645	0.000

(b)

i	treatment	R _n -G		λE		H		G		EF	
		mean	σ	mean	σ	mean	σ	mean	σ	mean	σ
1	top SR W	328.4	23.4	158.1	42.1	169.9	46.4	27.3	3.5	0.473	0.104
2	<top> SR W	330.2	10.4	158.8	41.9	171.1	43.9	27.3	2.7	0.473	0.104
3	top <SR> W	328.0	23.4	156.4	24.3	171.1	27.5	27.3	2.0	0.469	0.046
4	top SR <W>	328.4	23.4	156.6	32.9	171.3	38.8	27.2	3.7	0.467	0.088
5	<top> <SR> W	329.9	10.6	157.1	23.3	172.3	23.6	27.4	1.7	0.469	0.046
6	<top> SR <W>	330.1	10.3	157.4	32.7	172.5	35.8	27.2	3.0	0.467	0.088
7	top <SR> <W>	328.1	23.3	154.8	7.0	172.7	16.9	27.2	2.3	0.463	0.010
8	<top> <SR> <W>	330.1	0.0	155.5	0.0	174.6	0.0	27.4	0.0	0.462	0.000



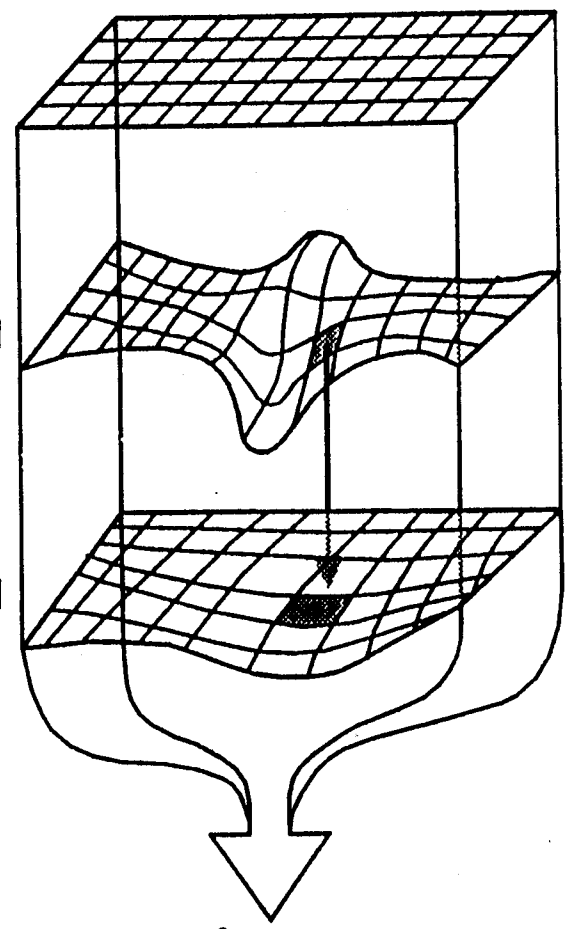
Area-Averaging

$$\langle Q \rangle_M = f \langle x_i \rangle$$

Flux $Q = f(x_i)$

Field of x_i

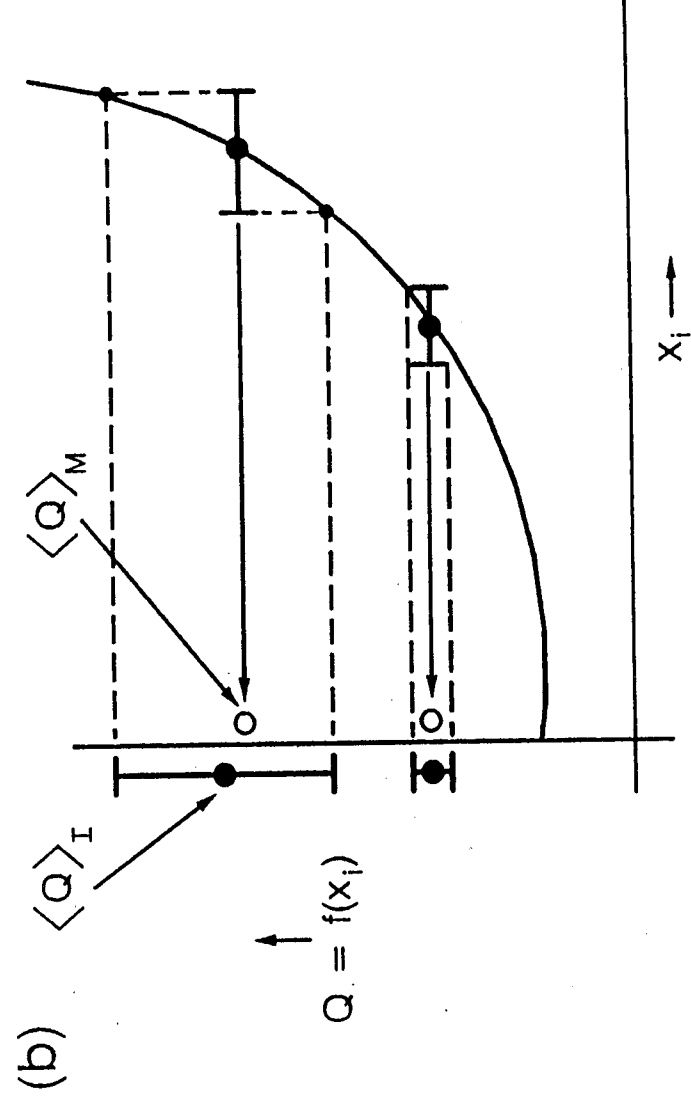
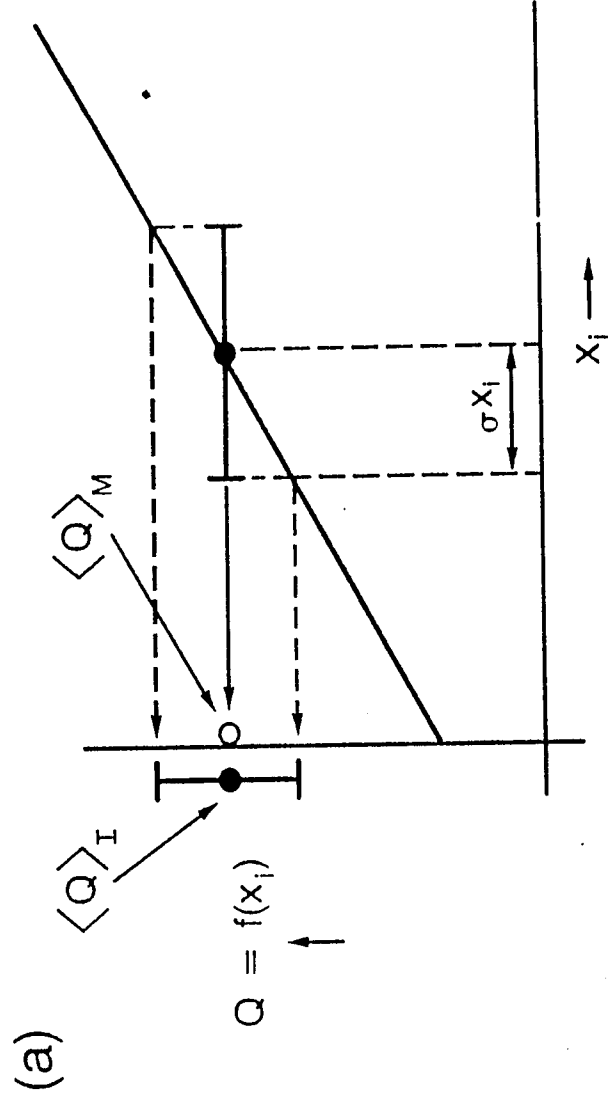
Domain A



$$\langle Q \rangle_I = \frac{1}{A} \oint_0^A f(x_i) da$$

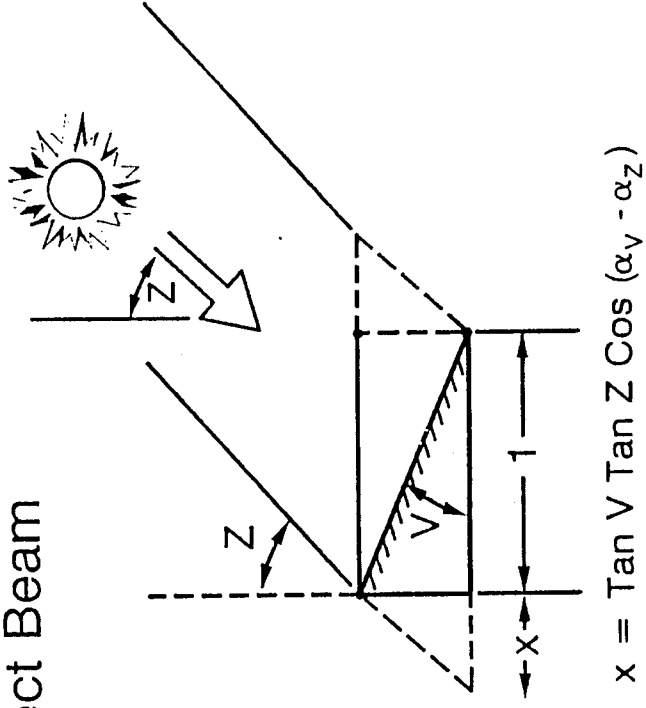
Area-Integration

Figure 2

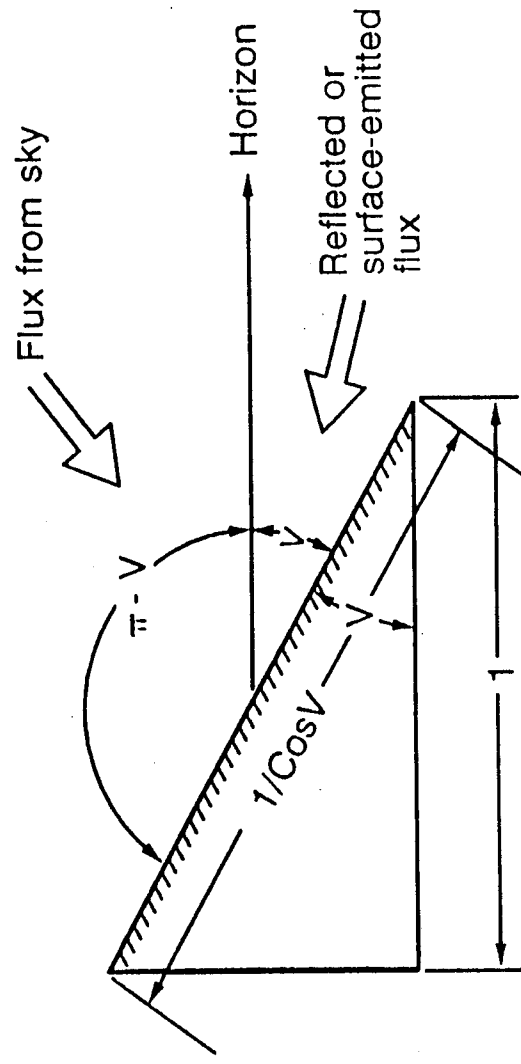


Figure

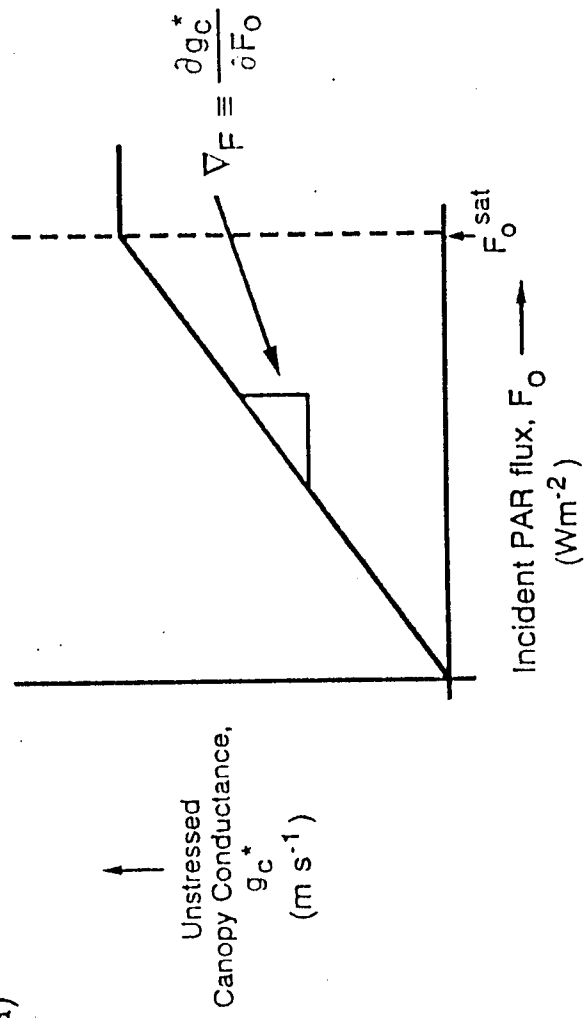
(a) Direct Beam



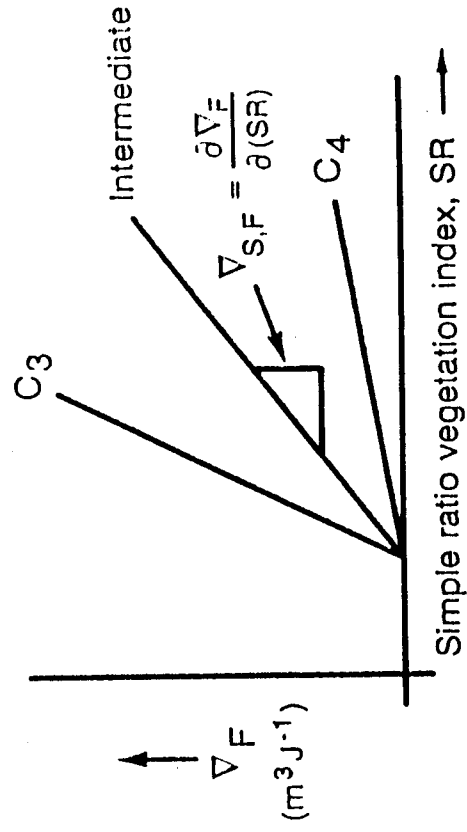
(b) Diffuse Radiation



(a)

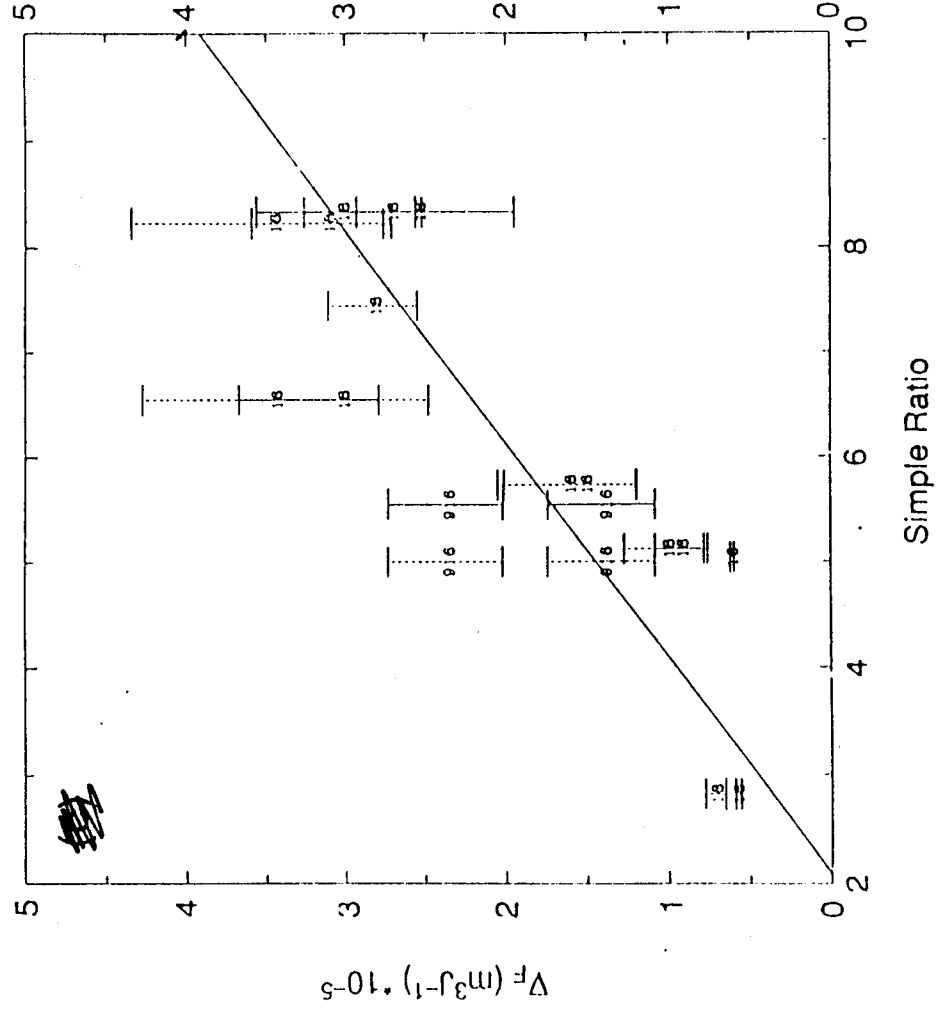


(b)

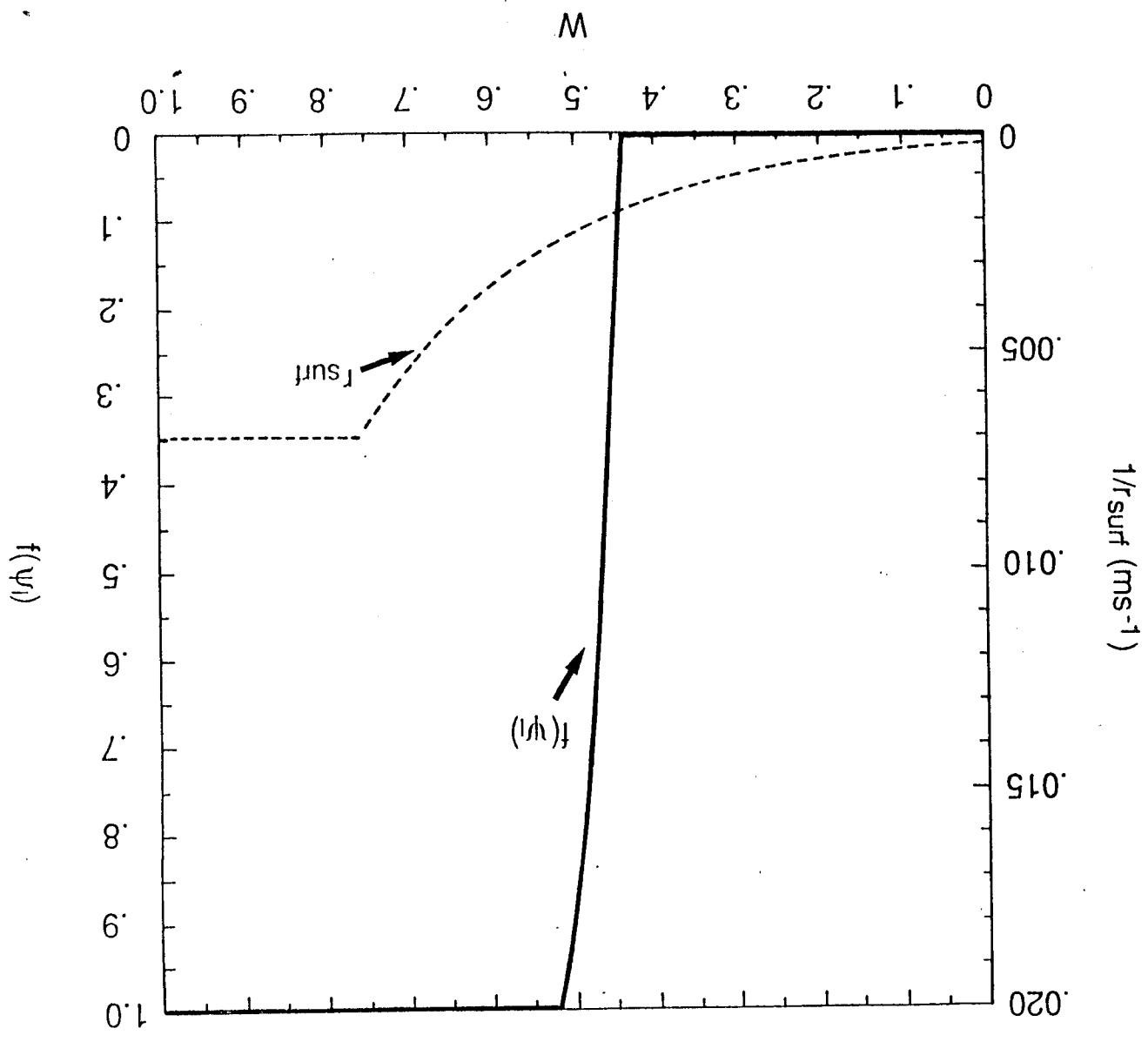


Figure

figure 6



Figure



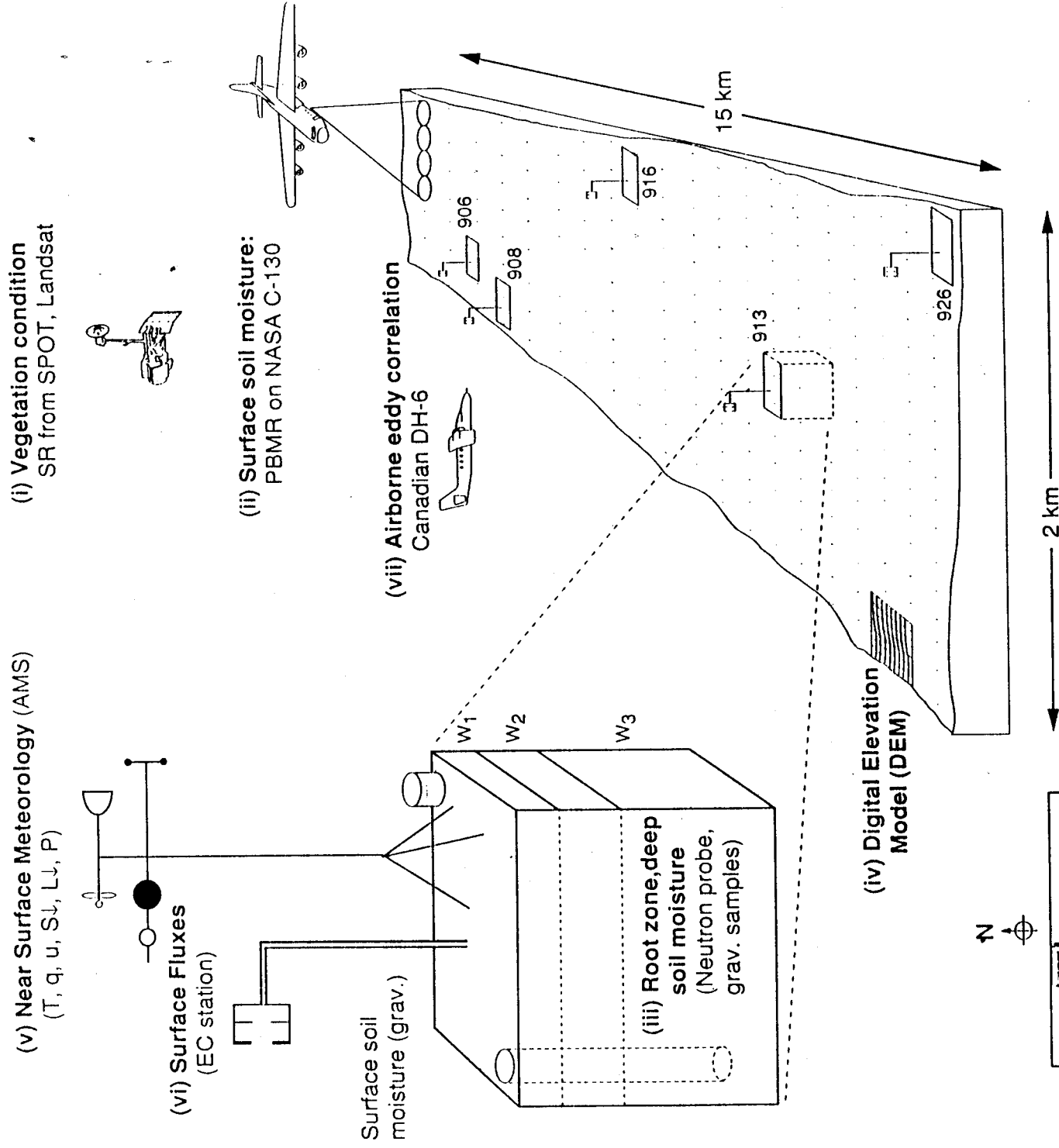


Figure 8

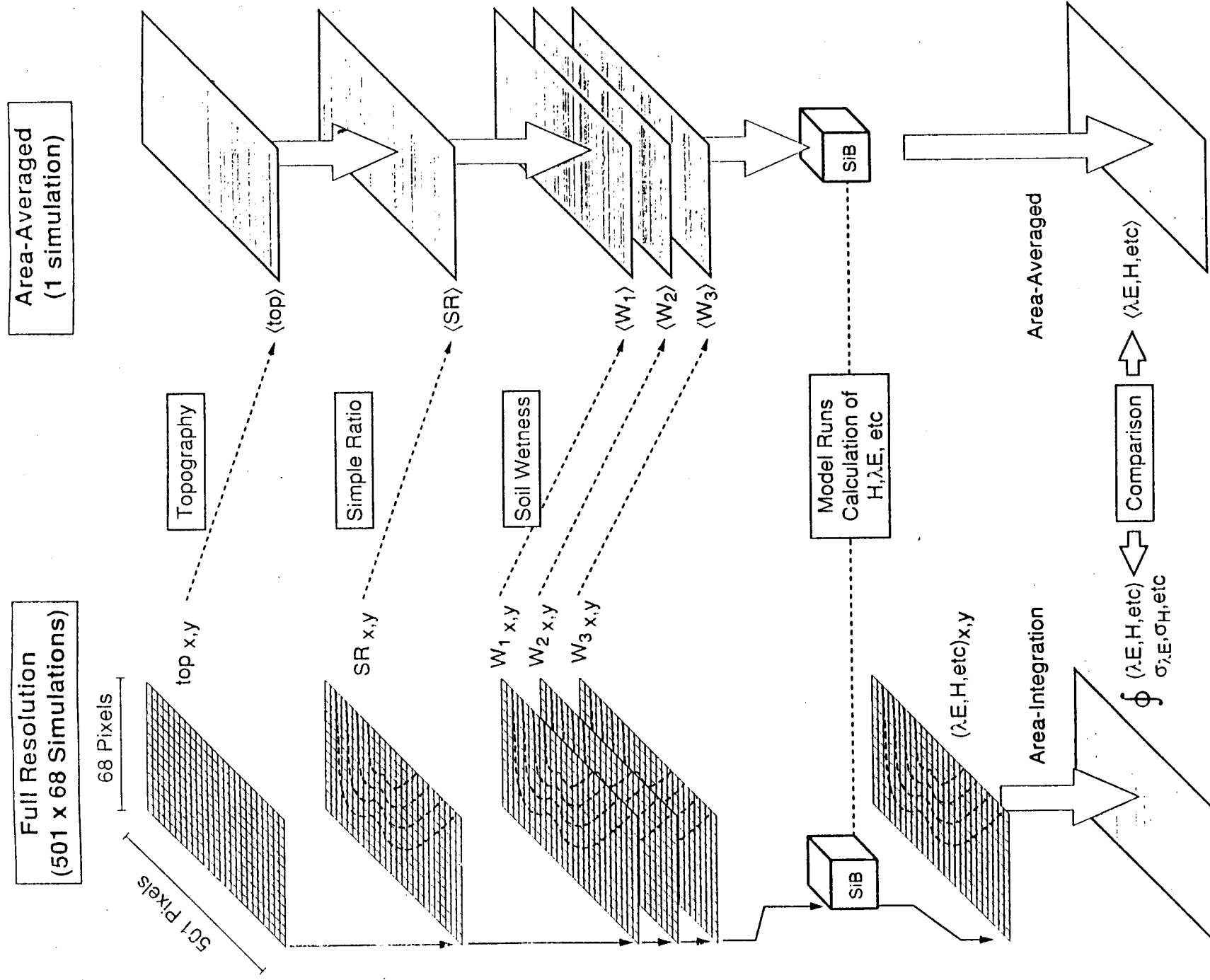
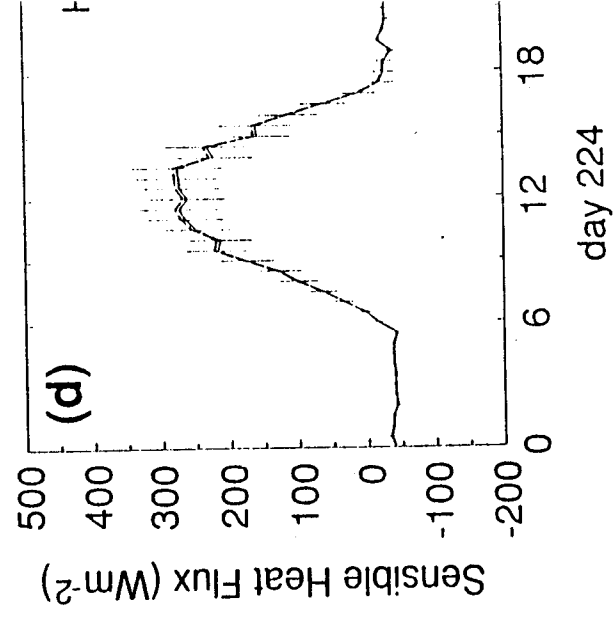
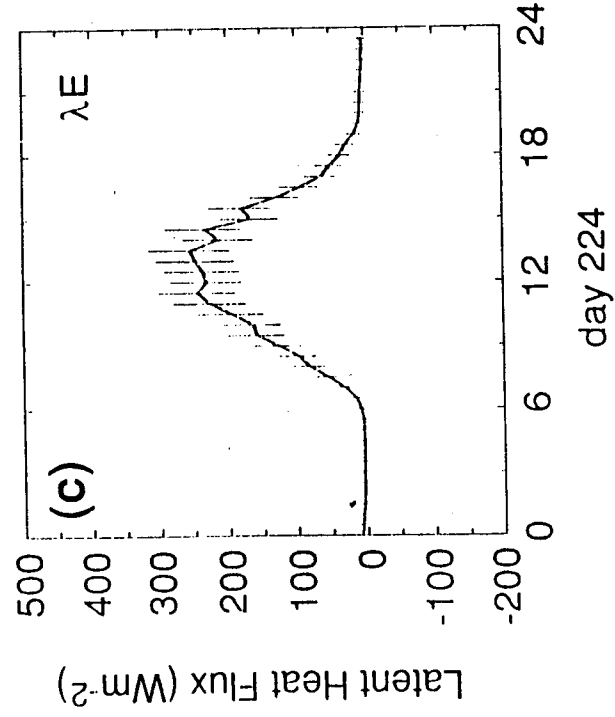
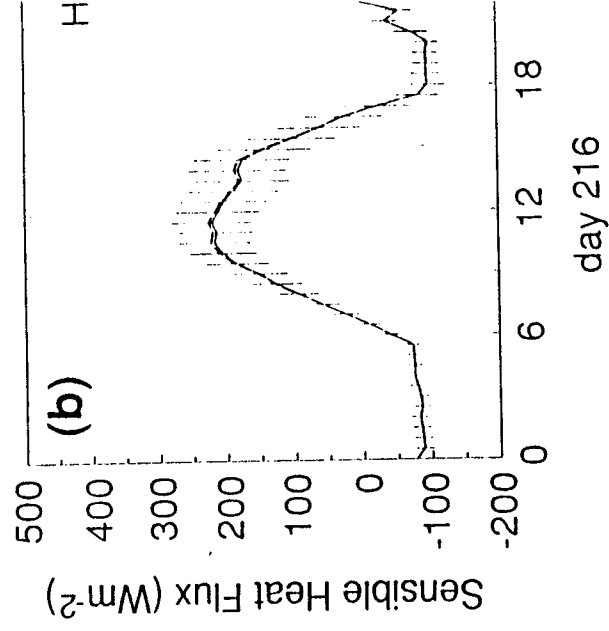
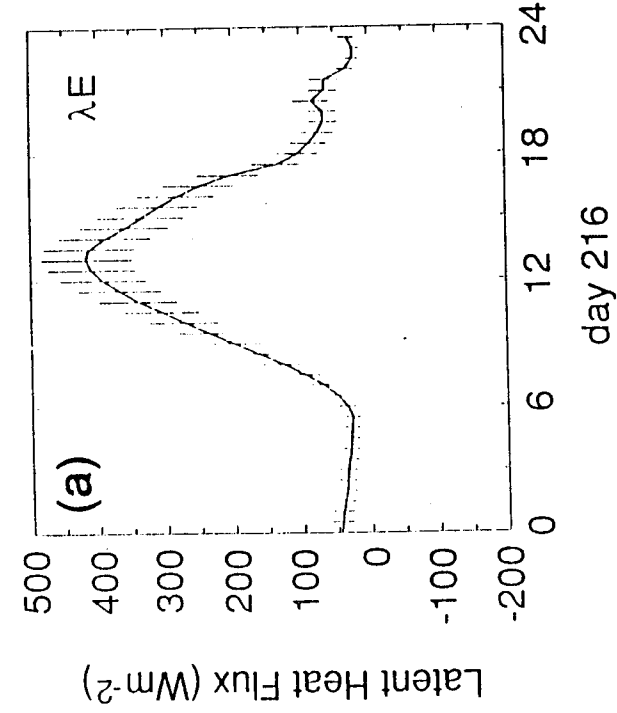
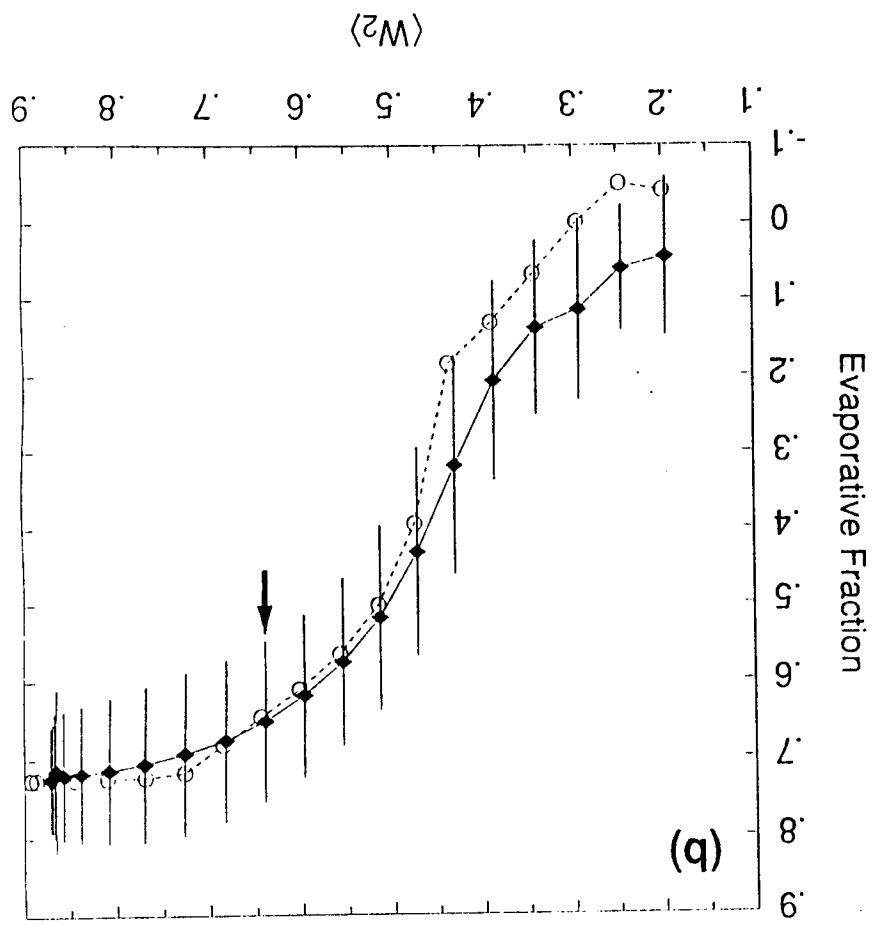
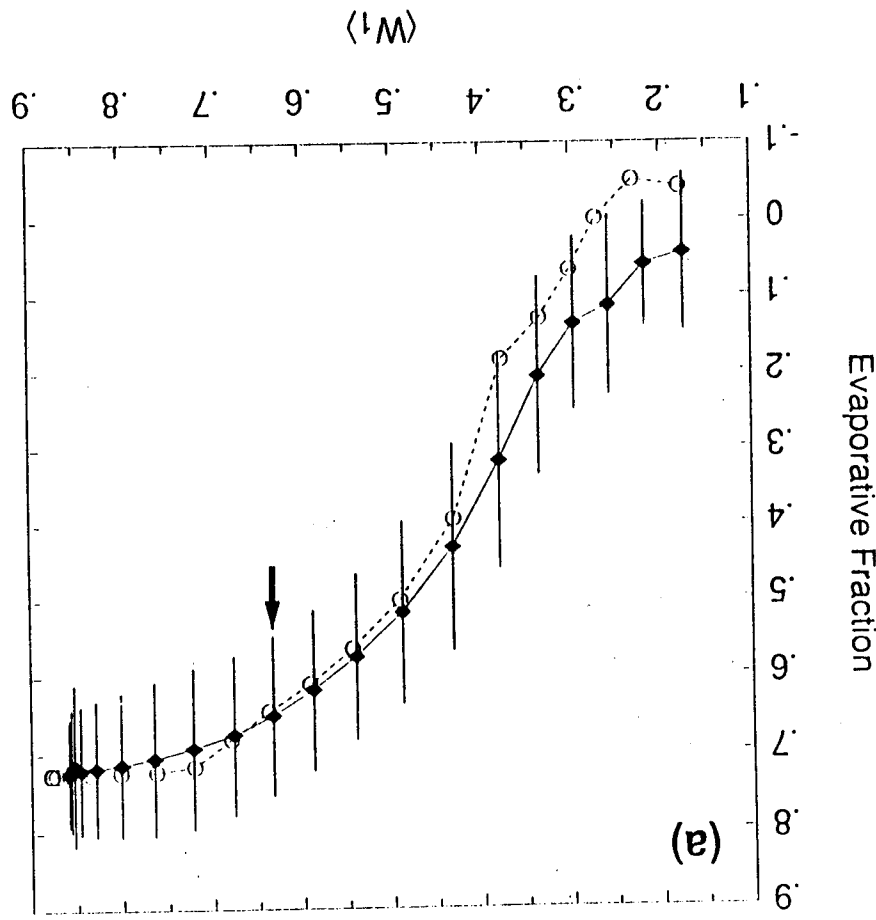


Figure 6



Figure

Figure 12



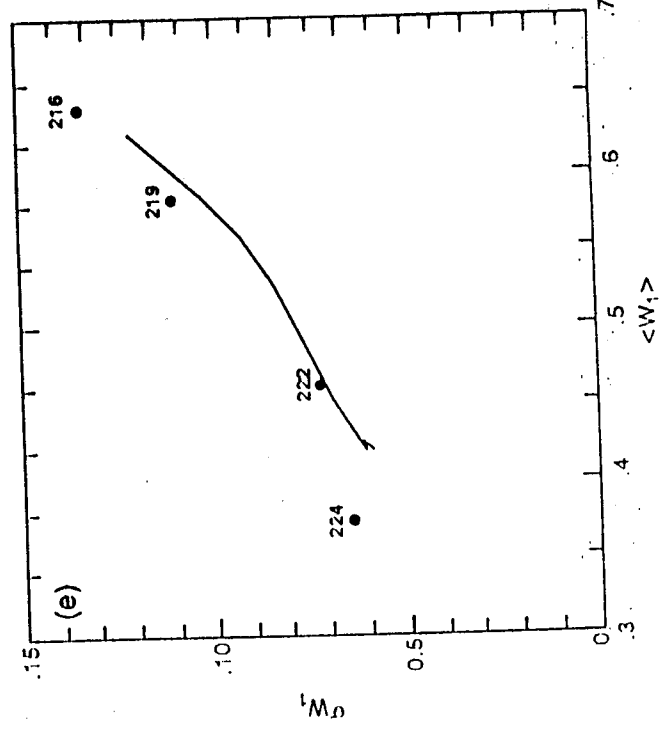
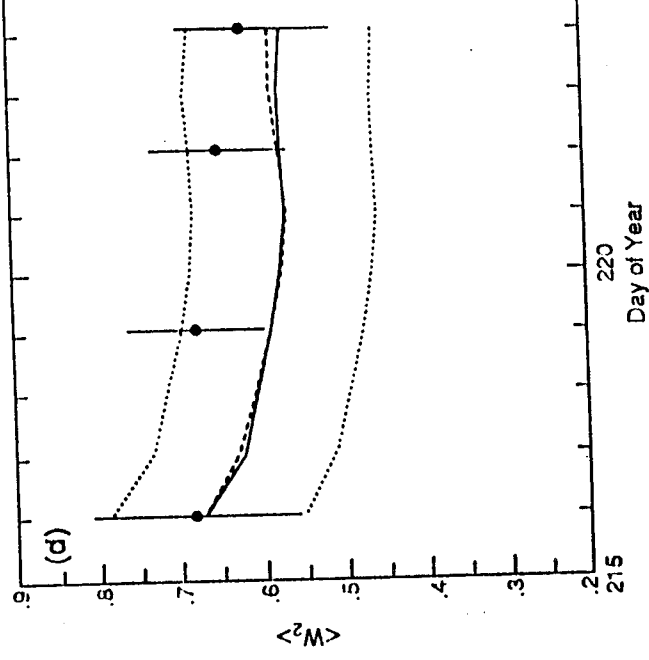
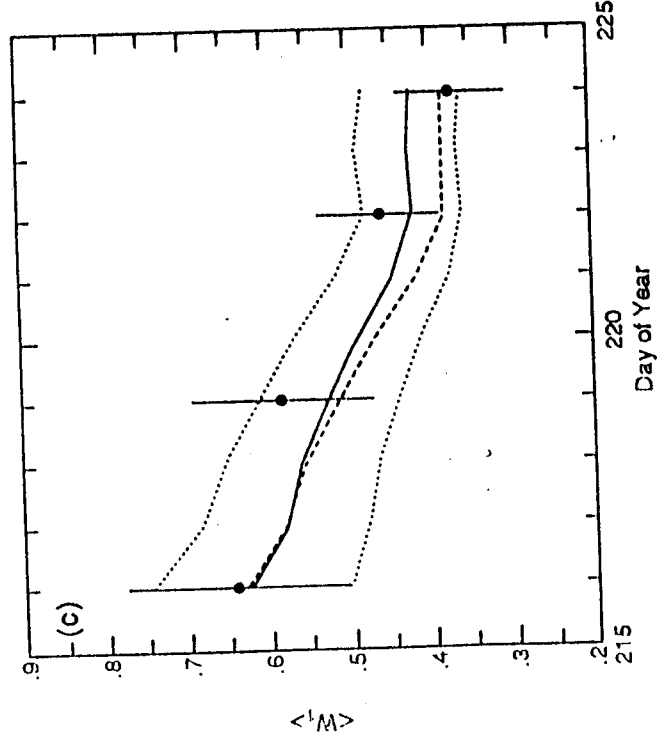
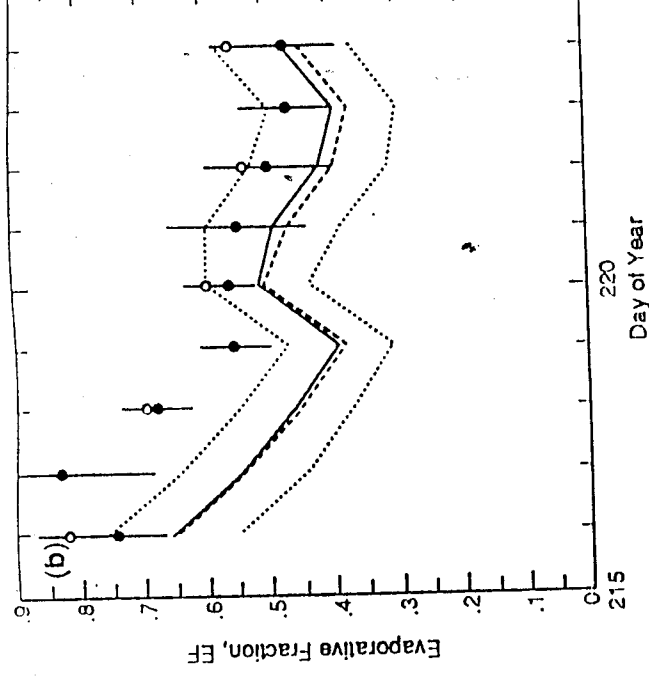
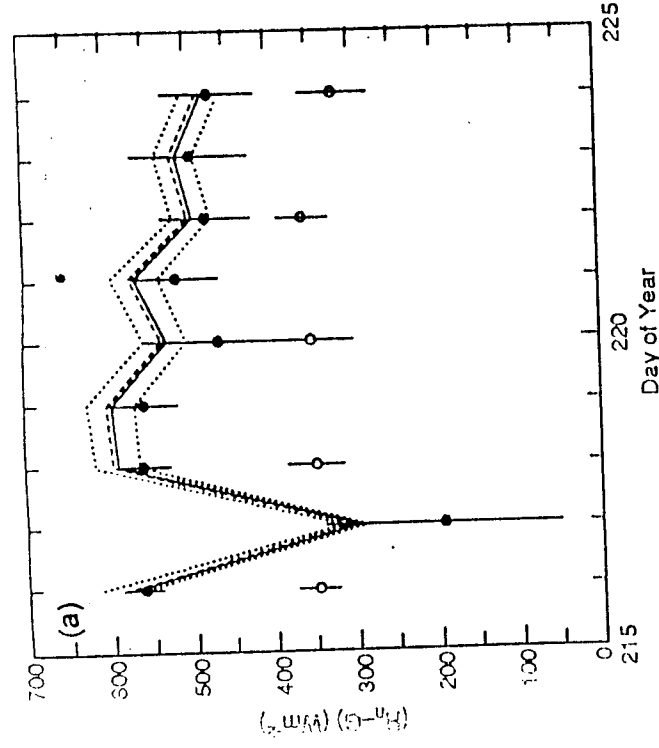


Figure 13

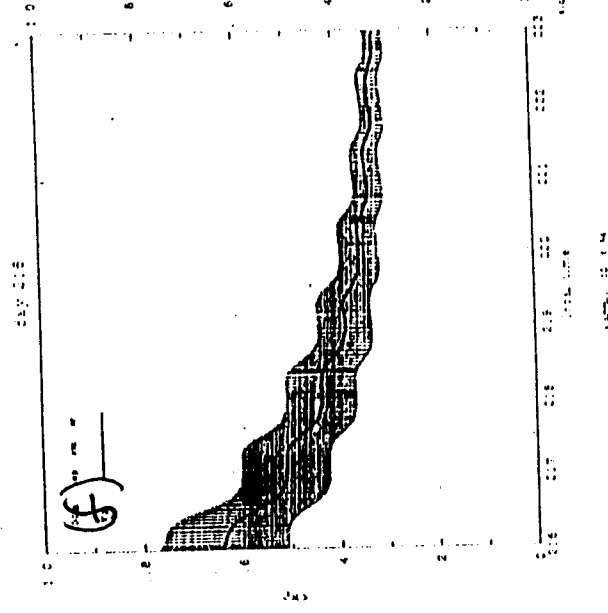
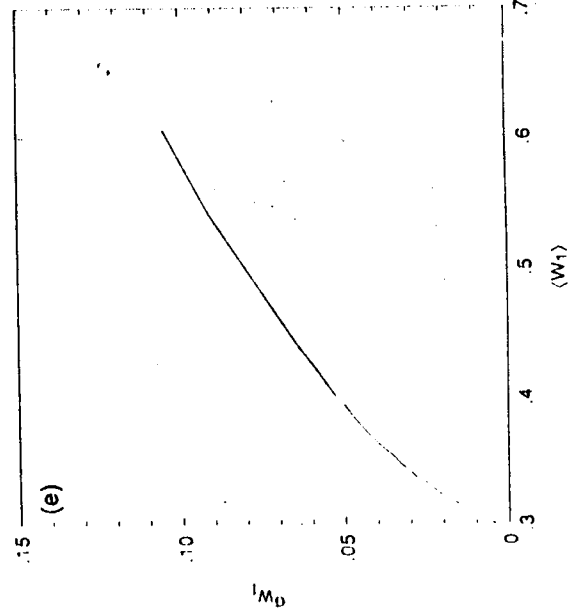
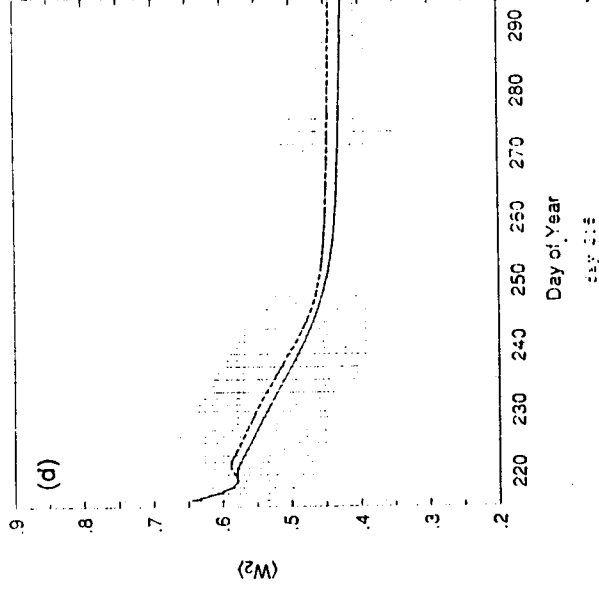
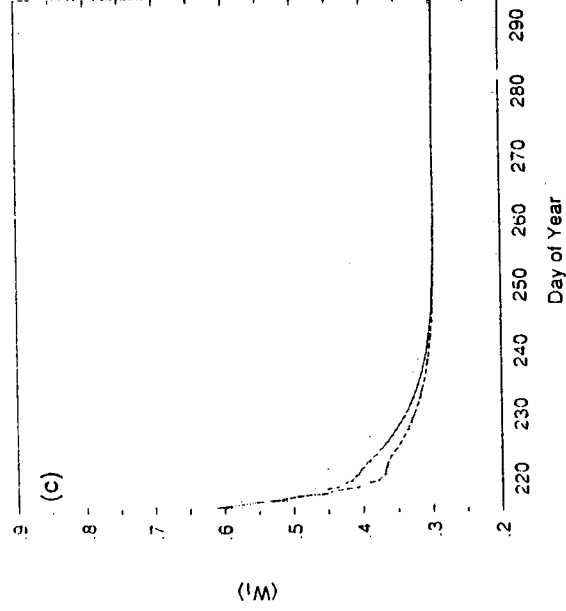
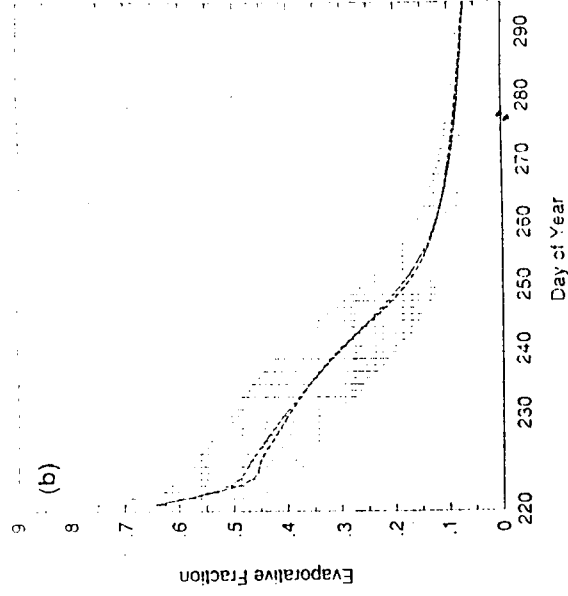
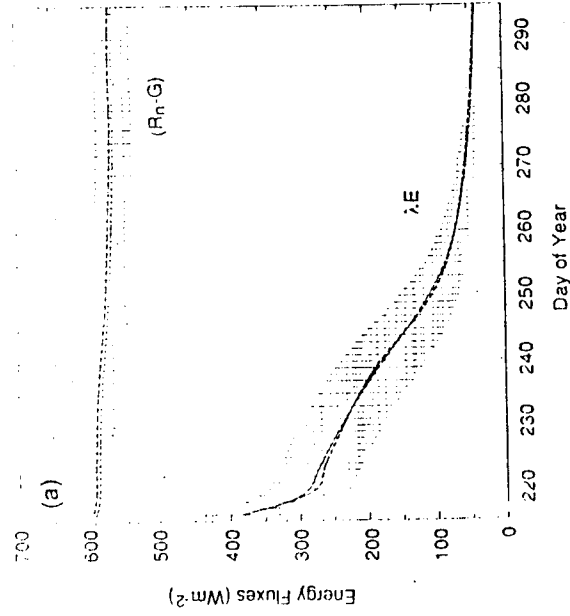


Figure 14

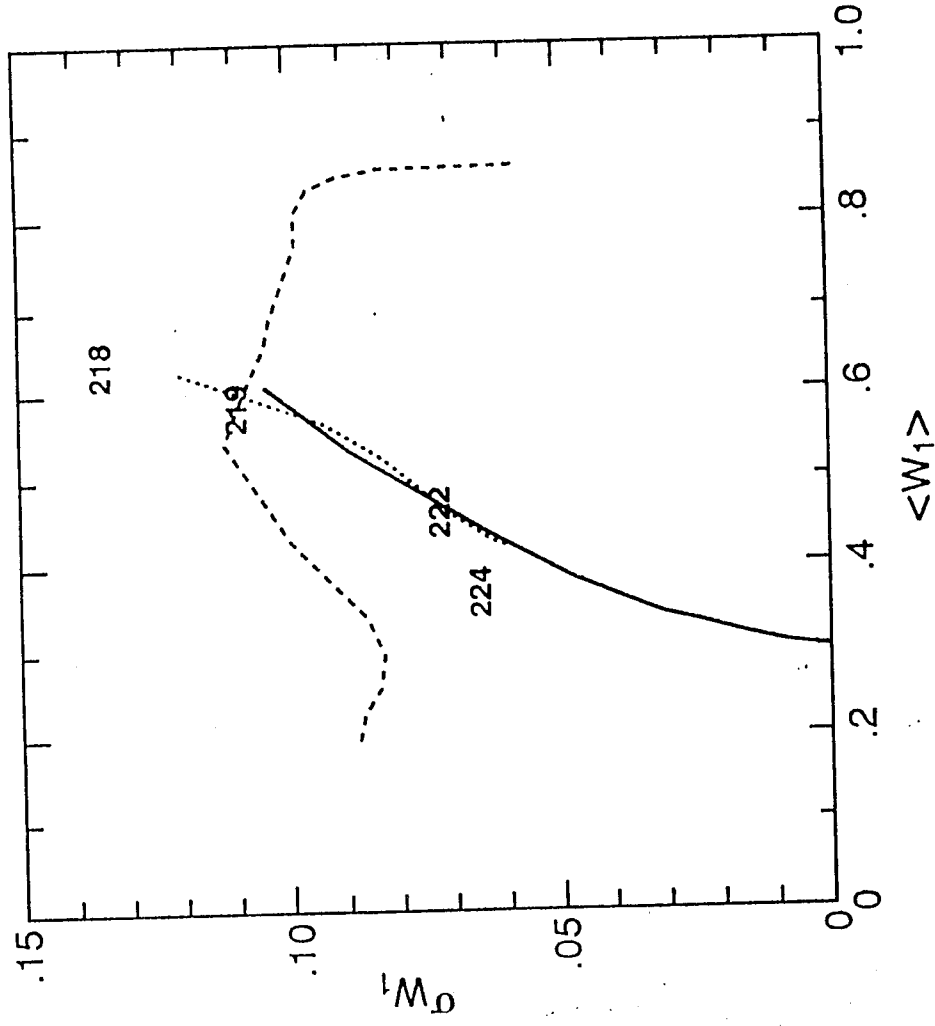


Figure 15

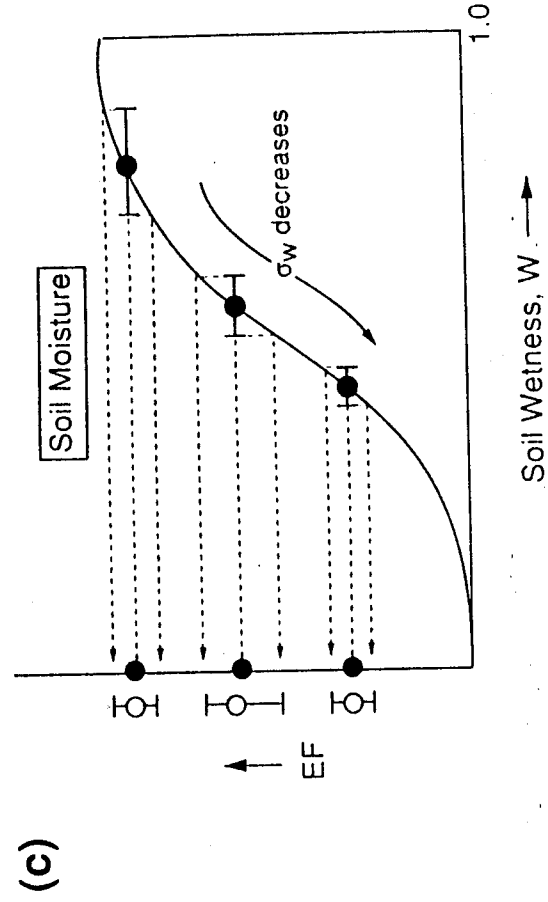
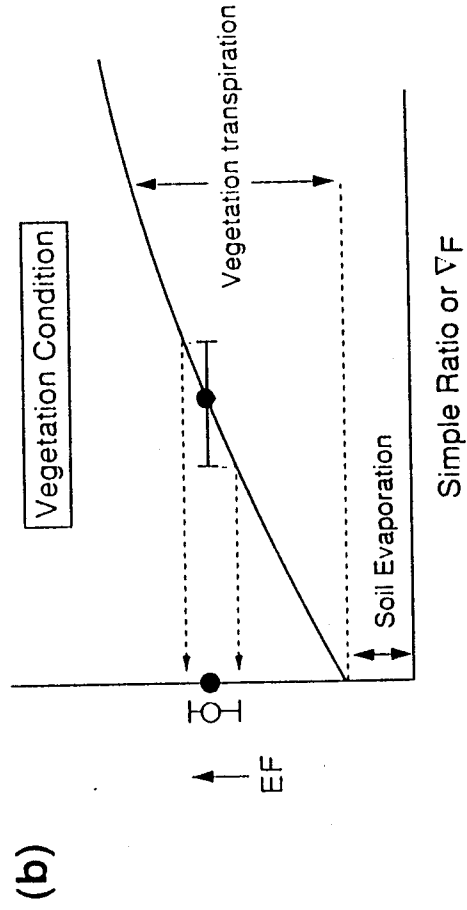
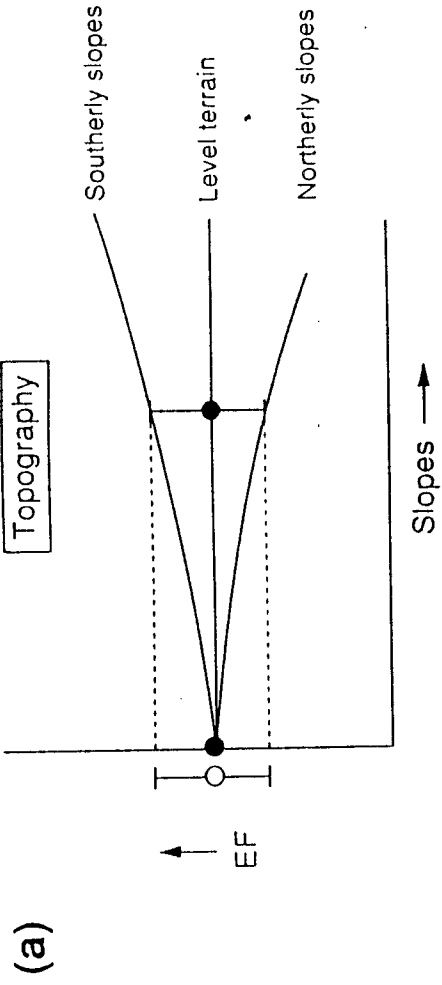


Figure 16

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Mapping daily surface meteorological variables
in complex terrain with MTCLIM-3D.

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Abstract

We present the logical and algorithmic framework of a numerical model MTCLIM-3D, which generates daily surfaces of maximum air temperature (Tmax), minimum air temperature (Tmin), dewpoint temperature (Tdew), precipitation (Prcp), and incoming shortwave radiation (Srad), at ground-surface level, over gridded terrain. We demonstrate the application of this model for one year (1989) over a grid with 2 km spacing covering 821,000 km² in the Columbia River Basin of the northwestern United States. Inputs include digital elevation values, daily observations of Tmax, Tmin, and Prcp from a network of surface observation sites. Interpolation over the grid is made sensitive to the horizontally heterogeneous distribution of stations by incorporating a smooth surface of station density, derived from the spatial convolution of a proximal polygon surface with a truncated gaussian filter kernel. Predictions are made sensitive to topographic variation through the spatially and temporally explicit diagnosis of the relationships to elevation of Tmax, Tmin, and Prcp, as well as through the influence of terrain height, slope, and aspect on predicted Srad. Elevation relationships are estimated from weighted linear regressions of temporally smoothed observations of Tmax, Tmin, and Prcp against elevation.

Predicted annual average Tmax and Tmin, derived from daily predictions, had standard errors of 1.0 and 1.7 °C, and biases of 0.01 and 0.00 °C, respectively. Predicted annual total precipitation, based on daily predictions, had a standard error c

25% of actual annual total, with a bias of 0.00%. R^2 values for regressions of predicted vs. observed annual average Tmax and Tmin (linear), and annual total Prcp (log:log) were 0.90, 0.69, and 0.84, respectively.

Introduction

An unresolved problem in regional hydroecological analysis has been availability of accurate meteorological drivers at appropriate spatial and temporal resolution. There are a number of sources of regular meteorological data, none of which in their original state are adequate for regional hydroecological simulations (Band et al 1991, Pierce et al 1993, Wigmosta et al 1994). Although most countries support a network of surface weather stations, these provide only point measures of conditions, and in complex topography these point samples may not translate at all to accurate areal representations of the local mesoclimate. Satellite observations can provide continuous temporal coverage (e.g. GOES), but with rather coarse spatial detail. GCM data can be used, but their coarse, 2-4° lat/long cell sizes cannot resolve the topography of most watersheds (Liston et al 1994, Kite et al 1994). Mesoscale meteorological models could be an option (eg RAMS - Pielke et al 1992; MM4 - Giorgi et al 1993), but these models compute dynamics of the entire vertical atmospheric profile, producing compute loads too high for routine multi-year climatological with fine spatial detail. The needs of hydroecological modeling have been identified in the IGBP BAHF Focus 4 project, the Weather Generator, and our model is a type of Weather Generator that is optimized for use with ecological modeling (IGBP 1993).

Our model, MTCLIM-3D, computes surface meteorology from standard daily surface weather station datasets, and produces the necessary variables for hydroecological models. (Running et al

1987, Glassy and Running 1994). A primary design criteria for MTCLIM-3D was to accurately represent the spatial variability of surface conditions in complex mountainous terrain at high spatial resolution, and provide continuous mapping for multiple year time periods. Topography then becomes an important variable for landscape aggregation. Our landscape modeling is much more efficient when topographically and microclimatically equivalent hillslope facets can be defined for hydroecological analysis (Band et al. 1993, Ford et al. 1994).

←
Simulations of ecosystem processes such as evapotranspiration, photosynthesis, respiration, and decomposition require meteorological inputs of air temperature, humidity, precipitation, and incident solar radiation. Spatial and temporal resolution of the hydroecological modeling domain are dictated in large part by the practicalities of data acquisition and computational loads. Our goal here is to provide meteorological inputs to a regional ecosystem process model which employs satellite derived estimates of leaf area index as a principal ecosystem descriptor (Running and Coughlan 1988). The availability of satellite data products helps to define the spatial resolution of these simulations. At regional scales, the NOAA/AVHRR satellite sensor, with a spatial resolution of about 1 km is used for estimation of leaf area for parameterization of terrestrial ecosystem models (Nemani et al 1993).

The temporal resolution of these applications is a compromise between a timestep which is short enough to provide accurate

•

predictions of ecosystem gas exchange processes, and long enough to allow multiple year simulations of carbon and nitrogen dynamics over many grid cells at an acceptable computational cost. The daily timestep is also standard for many sources of meteorological data. Archives of observations from U.S. National Weather Service Co-op stations and U.S. Soil Conservation Service SNOTEL stations provide daily point measurements distributed across the study area.

Over the past ten years we have developed a climate simulator sensitive to topography and that meets the need for accurate, high spatial resolution daily meteorological data (MT-CLIM, Running et al. 1987). The original MT-CLIM model did only single point-to-point extrapolations. Input consisted of daily observations of maximum and minimum temperatures, and precipitation, from 1-2 points, in addition to physical characteristics describing the base observation point(s) and the extrapolation point. Other ^{required} inputs ~~required~~ include; lapse rates for temperature and dew point, and base and site isohyets for extrapolation of precipitation. Output consists of daily predictions of temperatures, precipitation humidity, vapor pressure deficit, and incident solar radiation for the extrapolated point. Studies have demonstrated the usefulness of this original MT-CLIM logic for simulating meteorology for ecosystem process models on the scale of a single catchment (Band et al. 1991, Band et al. 1993, White and Running 1994).

We have now developed a model, MTCLIM-3D, that extends the MTCLIM logic by generating daily surfaces of meteorological variables over large regions, incorporating topographic databases and

observations from an unlimited number of arbitrarily spaced stations. Daily predictions of Tmax, Tmin, and Prcp for each cell over a gridded surface are generated by an interpolation algorithm, followed by calculation of Tdew and Srad by standard MT-CLIM algorithms. The interpolation logic also provides a spatially and temporally explicit diagnosis of the temperature and precipitation relationships with elevation, eliminating the need for the user to supply these parameters.

The MTCLIM-3D logic is presented here, along with examples of output from a 1989 simulation over a major river basin area in the northwestern United States. Cross-validation is used to generate estimates of the errors associated with the interpolated surfaces of temperature and precipitation. The radiation and humidity routines in MTCLIM-3D are similar to MT-CLIM (see Glassy and Running 1994), and their regional implementation will be described in a future paper.

Methods

Study Site Description

The region chosen for development and testing of this model is the Columbia River Basin of the Northwestern United States. This 820,924 km² area provides a testbed of incredible diversity, elevations ranging from 0 - 4500m, annual precipitation ranges from 10 - 300+ cm, annual average maximum temperatures ranging from 4.0 - 19.0°C, and a complete range of biome types from deserts to evergreen forests, grasslands and croplands. Hydrologically, the

Columbia River is the 2nd largest in the United States, and is at the center of ongoing controversy concerning natural resource management of hydroelectric power, salmon fisheries, forest harvesting and irrigation.

Model Description

Input requirements

The primary terrain input to MTCLIM-3D is a gridded surface of elevation, that is also used to derive surfaces of slope and aspect. This study used a digital elevation dataset for the conterminous United States with a horizontal resolution of 500 m and a vertical resolution of approximately 6 m, obtained from the U.S. Geological Survey EROS Data Center. Slope and aspect were derived from the 500 m data, and all terrain data was aggregated to a grid with 2km resolution. The interpolation logic requires a meteorological database with daily observations of Tmax, Tmin, and Prcp at each station in an arbitrarily spaced network. The stations need not lie within the bounds of the simulation's spatial domain it is desirable to have stations that lie outside the border of the simulation domain to help alleviate edge-effect errors. Missing values in the observation dataset are filled using the same algorithms described below for interpolations over the grid, and these values are flagged to avoid their use in the cross-validation error analysis. Stations with excessive missing data are dropped from the analysis.

For the simulations described here, any station with more than five consecutive missing values in any field or with more than 2

missing values for a given field over one year was dropped from the station database. These criteria resulted in the inclusion of 610 stations that recorded daily Tmax and Tmin (489 National Weather Service Co-op network (NWS) stations, and 121 Soil Conservation Service SNOTEL stations), and 821 stations that recorded daily Prcp (524 NWS and 297 SNOTEL stations). Figure 1 shows the gridded terrain and the locations of all observation points. Elevation, latitude, and longitude for each station are also required as inputs. Although the model presented here is applied to a uniformly gridded spatial domain, this logic is applicable to variable partitioned domains with only minor modifications.

Interpolation logic

Recent work by Daly et al (1994) has shown new methods can be used to interpolate fields of monthly or annual climatological precipitation over large and climatically diverse regions, and that those methods provide better estimates of annual climatological precipitation than do common geo-statistical methods based on kriging (Tabios and Salas 1985; Phillips et al 1992).

The interpolation logic employed by MTCLIM-3D is based on the generation, for each grid cell in the model domain, of a weighted list of stations that contribute to the prediction of meteorological fields at that cell. The weights associated with the stations in the list are determined by a spatial filtering process that gives greater weights to stations that are near the point of prediction, and that also modifies the diameter of the weighting kernel based on the density of stations near the point of

prediction. The spatial filtering consists of three steps: First, the spatial domain is partitioned into regions (referred to below as proximal polygons) that associate each cell in the grid with a single station, based on a nearness algorithm.' Next, a truncated gaussian filter kernel is convolved with the proximal polygon surface to generate a smooth surface of local station density. This station density surface is used to set the diameter of the spatial filter kernel employed in the third step. This final step does another spatial convolution that results in the generation, for each grid-cell, of both a list of stations to be included in the interpolation and a list of weights associated with those stations. These three steps are detailed below.

Generation of Proximal Polygon Surface

The first step in the interpolation process is the generation of a set of proximal polygons which partition the spatial domain such that: Given a set of points, s (the stations in this case) each member s_i of s is associated with a polygon which is the locus of points (x, y, z) in the spatial domain that are nearer to s_i than to any other member of s (after Preparata and Shamos, 1985). The simplest criteria for nearness on a topographic surface is:

$$D_{p,i} = \sqrt{(x_p - x_i)^2 + (y_p - y_i)^2 + (z_p - z_i)^2} \quad \text{Eq 1}$$

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Where $D_{p,i}$ is the distance from an arbitrary point (x_p, y_p, z_p) to a point (x_i, y_i, z_i) which is a member s_i of the set s .

Neglecting the contribution from the z-component of $D_{p,i}$, this criteria is the foundation of the Thiessen method for the spatial

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distribution of precipitation (Thiessen, 1911; Tabios and Salas 1985). The average density of stations in the NW U.S. database is about one station per 1100 km², resulting in contributions to $D_{p,i}$ that are much greater for the horizontal dimensions than for the vertical dimension. Consequently, the proximal polygon surfaces generated using this criteria are nearly indistinguishable from Thiessen polygon surfaces for the same region.

Generation of Station Density Surface

The goal of the interpolation scheme is to define a list of stations that are to be included in the daily predictions for each grid-cell in the domain. Since the distribution of stations is generally horizontally heterogeneous, some subregions of the domain will have considerably higher local densities of meteorological observations than others. There is a minimum standard for the number of stations that must be included in the interpolations at each point. At least two stations are required for a smooth horizontal interpolation using the methods described here, and at least six are required for regressions used for vertical extrapolations.

An undesirable result of this strategy is that, while data-sparse regions are supplied with sufficient information, very large numbers of stations are included in the interpolations for data-rich regions. Our goal was to have consistency in the number of stations included in the interpolation lists by incorporating a smooth surface of station density in the interpolation algorithm. The station density surface is generated by convolving the proximal

polygon surface with a spatial filter kernel of circular extent and fixed diameter. The kernel is defined on a regular grid with the same resolution as the simulation spatial domain, and a value, $V_{i,j}$, associated with each cell within the kernel is defined as:

$$V_{i,j} = 0 \quad \text{for: } D_{i,j} > R_{dk}$$

$$V_{i,j} = e^{-\frac{D_{i,j}^2 \alpha_{dk}}{R_{dk}^2}} - e^{-\alpha_{dk}} \quad \text{for: } D_{i,j} \leq R_{dk}$$

EQ 2

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where R_{dk} is the radius of the circular kernel, $D_{i,j}$ is the distance from the center of cell (i,j) to the center of the kernel grid, and α_{dk} is a shape parameter.

These equations define a truncated gaussian weighting function within the circular kernel, with the greatest weight located at the center of the kernel and decreasing radially outward to zero at a distance R_{dk} from the center of the kernel grid.

The kernel defined by R_{dk} and α_{dk} is spatially convolved with the proximal polygon surface to generate, at each cell in the spatial domain, the proportion of total kernel area and proportion of total kernel weight associated with each station, pa_s and pw_s respectively, defined as:

$$pa_s = \frac{\sum_{i,j=-R_{dk}}^{R_{dk}} M_{s(i,j)} M_{dk(i,j)}}{\sum_{i,j=-R_{dk}}^{R_{dk}} M_{dk(i,j)}} \quad pw_s = \frac{\sum_{i,j=-R_{dk}}^{R_{dk}} M_{s(i,j)} V_{dk(i,j)}}{\sum_{i,j=-R_{dk}}^{R_{dk}} V_{dk(i,j)}} \quad \text{EQ 3}$$

Where s is the station index, (i,j) are spatial coordinates within the filter kernel, $M_s(i,j)$ is a mask associated with the proximal polygon for station s , having a value of 1 inside the polygon and 0 elsewhere, $M_{dk}(i,j)$ is a mask associated with the spatial kernel, having a value of 1 for cells with $V_{dk}(i,j) > 0$, and 0 elsewhere.

Based on these intermediate variables, the station density at each point is defined as:

$$SD = \frac{1}{ka \left(\sum_{s=1}^{ns} pw_s pa_s \right)} \quad \text{EQ 4}$$

Where SD is the station density (stations/km²), ka is the kernel area (km²) with non-zero weight, s is the station index, and ns is the number of stations for which some portion of the stations' proximal polygons intersect the non-zero weighted region of the spatial kernel.

Generation of Weighted Station Lists

In the final step of the interpolation process, a new spatial convolution kernel is generated, where the radius varies as a function of the station density surface as follows:

$$R_{1k(i,j)} = \frac{\sqrt{\frac{NS}{SD(i,j)}}}{C_1} \quad \text{EQ 5}$$

Where $R_{1k(i,j)}$ is the radius (km) of the station list filter kernel at point (i,j) in the simulation spatial domain, NS is the average number of stations to include in the station list at any point, specified by the user, and C_1 is a unitless parameter to correct for the influence of kernel truncation on estimated density. C_1 is determined empirically, and is generally assigned a value between 2.0 and 3.0.

The user specifies a shape parameter, α_{1k} , and the station list convolution kernel is then defined as in Eq.2 above. In practice, we set $\alpha_{dk} = \alpha_{1k}$. The station list for a given cell has one entry for each station which has any part of its proximal polygon within the non-zero weighted region of the station list kernel centered

over that cell. It is therefore not required for inclusion that the station be located within the bounds of the kernel; only some portion of its proximal polygon must do so for the station to be included in the list. The weight assigned to each station, w_s , is defined in the same way as the proportional weights, pa_s , in Eq. 3 above. Since it is common at the edges of the domain to have a part of the kernel extending outside the domain, weights derived in this way are summed over all included stations at a point and normalized to that sum, forcing the normalized sum of weights for all stations at a given point to 1.0. The weighted list of included stations is generated one time for each grid cell; the same list is used for predictions at each timestep.

Histograms of station-list length over all cells in the study region were plotted for interpolations based on: (1) list kernel radii which have been modified for station density according to Eq 5, and (2) a constant list kernel radius set equal to the average of the modified radii, to check for the effect of station density on the interpolation logic.

Estimation of Elevation Relationships

We recognize that there are many factors influencing the observed daily values of Tmax, Tmin, and Prcp at a given point. These include elevation, exposure, patterns of synoptic and surface wind flow, atmospheric stability, cloudiness, and other microterrain-influenced processes such as cold air drainage, frost formation and shading. In addition, there are many potential sources of human and instrumental error in the observations.

databases. The interpolation logic described above addresses only one aspect of the total variation in these observations, the horizontal variations of the meteorological fields on any given day. The spatial averaging techniques described above help dampen the influence of observational error and unresolved physical processes on these horizontal distributions. In addition, they provide us with a framework within which we can assess the spatial and temporal variation of our primary variables with elevation. In order to reduce the variability imposed on these relationships by extraneous factors, we incorporate a temporal smoothing of the observed fields in addition to the spatial smoothing accomplished by the station weighting algorithms.

A linearly ramped window smoothing algorithm is applied to the time series of observed values at each point, and the smoothed values for each day are used to generate the regression statistics with elevation. This smoothed data is used only in the generation of the elevation relationships; daily predictions are based on the unsmoothed observations. The width, in days, of the smoothing windows for temperature (both $T_{\max}$ and $T_{\min}$) and precipitation regressions are denoted W_{ts} and W_{ps} , respectively.

For each grid cell, maximum and minimum temperature lapse rates are independently estimated for each day of simulation. These estimates are based on a weighted least squares regression of the smoothed observations of $T_{\max}$ and $T_{\min}$ for the stations in a grid cell's weighted list. At each day, there is one point in these weighted regressions for each unique pairing of stations in

the cell's list, giving the number of points in each regression as:

$$N = \frac{n_{list}^2 - n_{list}}{2} \quad \text{EQ 6}$$

where N is the number of points in the regression, and n_{list} is the number of stations in the weighted list at the cell in question.

The weight associated with each regression point is defined as the product of the two weights associated with the stations in a pair. The independent variable is defined as the difference in elevation between the stations in a pair, and the dependent variable is defined as the difference in the smoothed daily values of either Tmax or Tmin between the stations in a pair, giving a regressor equation of the form:

$$T_1 - T_2 = \beta_0 + \beta_1(Z_1 - Z_2) \quad \text{EQ 7}$$

where T_1 and T_2 are the smoothed observations of either Tmax or Tmin at two points making up a pair, β_0 and β_1 are the y-intercept and slope, respectively, of the regression line, and Z_1 and Z_2 are the elevations of the two points.

Precipitation relationships with elevation are also estimated for each grid-cell, for each day, but the form of these regression is somewhat different than for the temperature relationships. We use a normalized difference algorithm to represent the between station variation in time-smoothed precipitation amounts, giving regression equation of the form:

$$\left(\frac{P_1 - P_2}{P_1 + P_2} \right) = \beta_0 + \beta_1(Z_1 - Z_2) \quad \text{EQ 8}$$

where P_1 and P_2 are the smoothed values of daily precipitation at two points making up a pair, β_0 and β_1 are the y-intercept and slope, respectively, of the regression line, and Z_1 and Z_2 are the elevations of the two points.

Since time-smoothed values of daily precipitation are not quantitatively comparable to the unsmoothed data used in the prediction algorithm, we require a regression algorithm that is independent of the magnitude of the dependent variable. By normalizing the independent variable for each point in the regression to the sum of the daily smoothed value for the two stations in a pair, β_0 and β_1 lose their dependence on the absolute precipitation amount. This formulation has the numerical advantage over both ratio and semi-log formulations of being bounded (-1.0 to +1.0) over the range of values for P_1 and P_2 (Figure 2).

Prediction of Temperatures and Precipitation

The prediction equation for both Tmax and Tmin at a particular location-day takes the form:

$$T_p = \sum_{s=1}^{n_{list}} W_s (T_s + \beta_0 + \beta_1(Z_p - Z_s))$$

EQ 9

where T_p is the predicted temperature, s is the index for stations in the station list, n_{list} is the number of stations in the list, W_s is the weight associated with station s , T_s is the observed temperature for station s , β_0 and β_1 are the regression parameters defined above, Z_p is the elevation of the point of prediction, and Z_s is the elevation of station s .

The prediction of daily surfaces of precipitation requires a more elaborate treatment (Chua and Bras 1982; Hevesi et al 1992, Phillips et al 1992). The sampling density afforded by the station network is much too coarse to resolve the detailed spatial patterns of precipitation occurrence. This lack of spatial detail is somewhat offset by the use of a daily timestep, since over the

course of one day a wintertime frontal precipitation event will be recorded over a wide range. Warm season convective precipitation events, although of generally shorter duration, will likewise show more spatially coherent patterns when integrated over a day than they would at any instant. In general a simple prediction of the form used to estimate temperatures is inappropriate for the diagnosis of precipitation, since the weight contributed by stations that receive no precipitation for a day combined with the weight of stations that recorded precipitation tends to result in the over-prediction of event frequency and the under-prediction of event intensity.

Based on the assumption that over one day there is some level of spatial coherence in the precipitation event frequency, we propose a diagnostic, the precipitation trigger statistic (PT) which is used to make a binomial prediction of daily precipitation occurrence at a point. We define this statistic as the sum of the weights of the stations in a cell's weighted list that record an precipitation for the day. If that sum exceeds a critical value (PT_{crit}), a precipitation occurrence for the day is predicted and we continue by generating a prediction of the event intensity otherwise we predict a dry day. If all stations in a cell's list record precipitation for a day then $PT = 1.0$, if no stations record precipitation then $PT = 0.0$. Values of PT_{crit} close to 1.0 result in relatively few predicted events, whereas values close to 0.0 result in many predicted events. At present we parameterize PT_{crit} once for the entire spatial and temporal domain, based on a cross-validation

analysis of predicted vs. observed event frequencies.

If, for a given point on a given day, $PT > PT_{crit}$, a precipitation event intensity prediction is generated for that point. The incorporation of the PT statistic helps correct for the tendency to over-estimate event frequency, but the tendency to under-estimate event intensity must still be addressed. By incorporating in the daily predictions only those stations recording precipitation for the day, we generate event intensities which are more realistic than if all stations are included. The weights for the new list of included stations are renormalized to sum to 1.0, and the final event intensity prediction takes the form:

$$P_p = \sum_{s=1}^{n_{wetlist}} W_s P_s \left(\frac{1 + f}{1 - f} \right) \quad \text{EQ 10}$$

$$f = \beta_0 + \beta_1(Z_p - Z_s)$$

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down

where P_s is the predicted precipitation amount, s is the variable indexing the stations in the station list that recorded precipitation for the day, $n_{wetlist}$ is the number of such stations, W_s is the renormalized weight associated with station s , P_s is the observed (unsmoothed) precipitation at station s , β_0 and β_1 are the y -intercept and slope of the regression line.

The fraction on the right side of Eq. 8 expresses the conversion from the normalized regression parameters to predictions of actual precipitation amounts. For predictions at the highest grid-cells in mountainous terrain, the lack of nearby high-elevation observations can lead to values of f that are outside the range (-1.0 to +1.0) of valid inputs, as defined by the form of the dependent variable

in the Prcp vs. elevation regression equation (Eq. 8). We therefore set an upper limit on acceptable absolute values of f , denoted $f_{\max}$, where $1.0 > f_{\max} > 0.0$. The closeness of this parameter to 0.0 serves as a quantitative expression of the lack of confidence in the extension to mountain-tops of regression relationships developed over mostly low-elevation observations.

Results

Parameters and Interpolation

The parameter values used for the NW U.S. test case are as follows: radius of station density kernel, $R_{dk} = 101$ km; gaussian shape parameter for station density kernel, $\alpha_{dk} = 3.0$; average number of stations in station lists, $NS = 18$; station-list kernel radius correction, $C_1 = 2.7$; smoothing window widths for temperature and precipitation, W_{ts} and $W_{ps} = 11$ and 31 days, respectively; precipitation trigger criteria, $PT_{crit} = 0.5$; and precipitation regression truncation, $f_{\max} = 0.6$. The value for $f_{\max}$ was set determined based on the distributions of terrain height and station elevation in the vicinity of the highest terrain. Precipitation regression truncation was in effect over a very limited subset of the study area, about 800 km², or 0.1% of the total area, but it has a very strong effect on the precipitation predictions in that subregion. The average radius of the station list convolution kernel over the study area was 63 km, and the average number of stations per list was 17.9.

Elevation Relationships

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Values of β , diagnosed for temperature and precipitation from the regressions defined by Eqs. 7 and 8, respectively, were averaged over all cells in the study region for each day, and the resulting time series, illustrating the intra-annual variation in these spatially averaged relationships, are plotted for Tmax, Tmin, and Prcp (Figure 3). The November - February winter period shows a common weaker winter lapse rate, -2 - -6 °C/km for Tmax, the lapse rate the remainder of the year being closer to -8 °C/km. Tmin and precipitation elevational relationships also have a clear winter versus summer seasonality (Karl et al 1990).

Predicted Meteorological Fields

Daily predictions of Tmax, Tmin, and Prcp for 1989 were made over the study region, based on the weighted station lists and Eqs. 9 and 10. For brevity, only the annual average Tmax and Tmin predictions, and the annual total of the Prcp predictions, are shown in Figures 4 and 5, respectively.

Table 1 lists the results of a cross-validation analysis of predicted vs. observed Tmax, Tmin, and Prcp at the station locations. Comparisons of the observed and predicted annual total precipitation were performed on the log transformed values, since both observations and predictions are log-normally distributed. This also allows the expression of errors in terms of % of predicted values, which, given the large range of total annual prcp, is more informative than the expression of errors in units of depth. Scatter-plots corresponding to the regression statistics in Table 1 for predicted vs. observed annual average Tmax and Tmin and

annual total Prcp are shown in Figures 6 and 7. These cross-validation statistics show a low standard error, and almost no bias in our model outputs.

Discussion

The MTCLIM-3D model provides us with our first opportunity to perform hydroecological process simulations at a regional scale with accurate meteorological input surfaces derived from large databases of daily observations. The algorithms employed allow these surfaces to be generated rapidly (runs illustrated here took 8 hours running on an IBM RS6000 Model 350 workstation). The list of parameters required by these algorithms is short, making the application of these methods to a new model domain relatively easy and continental applications possible (Dolph and Marks 1992; Liston et al, 1994). We are also exploring the potential for incorporating surface temperature estimates from NOAA/AVHRR data into these algorithms as a way to constrain the regressions of observed variables against elevation.

We are currently exploring the effects of varying spatial resolution of the topographic input layer on the model behavior. Daly et. al (1994) report that for annual and monthly long-term average precipitation, a smoothed digital elevation map provides better estimates of the effective climatological elevation of an observation point than the actual elevation, and the same may be true for daily observations of precipitation and temperature. The degree to which complex landscapes can be aggregated for

microclimate mapping will greatly influence the computational efficiency of these algorithms for larger scale applications.

MTCLIM-3D can serve as a general tool for many applications in which daily surfaces of meteorological variables over complex terrain are useful, for example in the assessment of ecosystem disturbance regimes or land-management scenarios. The logical framework presented here can be configured to accomodate a wide range of input densities and spatial resolutions, and can provide both interpolated surfaces of meteorological variables as well as surfaces of the prediction statistics associated with those variables.

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Table 1

Cross-validation error estimates for observed vs. predicted variables, averaged over all stations in the Columbia Basin study area. The cross-validation was done by sequentially removing individual station points from each proximal polygon, and recalculating the Tmax, Tmin, and Prcp values. See Figures 6 and 7 for plotted data.

<u>Variable</u>	<u>R²</u>	<u>Std error</u>	<u>Bias</u>
Tmax	0.903	1.03 °C	-0.011 °C
Tmin	0.694	1.72 °C	0.000 °C
ln(Prcp)	0.844	25.0%	0.000%

Figure Captions

Figure 1. Digital elevation (m) of the 821,000km² Columbia River Basin study area in the Northwestern U.S., with surface weather observation stations marked.

Figure 2. The Normalized Difference Precipitation Index used to smooth station (P1) to station (P2) variability in daily precipitation intensity. See Eq 8.

Figure 3. (a) Daily air temperature lapse rates ($^{\circ}\text{C}/\text{km}$) for Tmax and Tmin, averaged for the entire Columbia Basin, for 1989. (b) Spatially-averaged relation of daily precipitation to elevation (cm/km).

Figure 4. (a) Annual average Tmax over the Columbia Basin study region. (b) Annual average Tmin over study region.

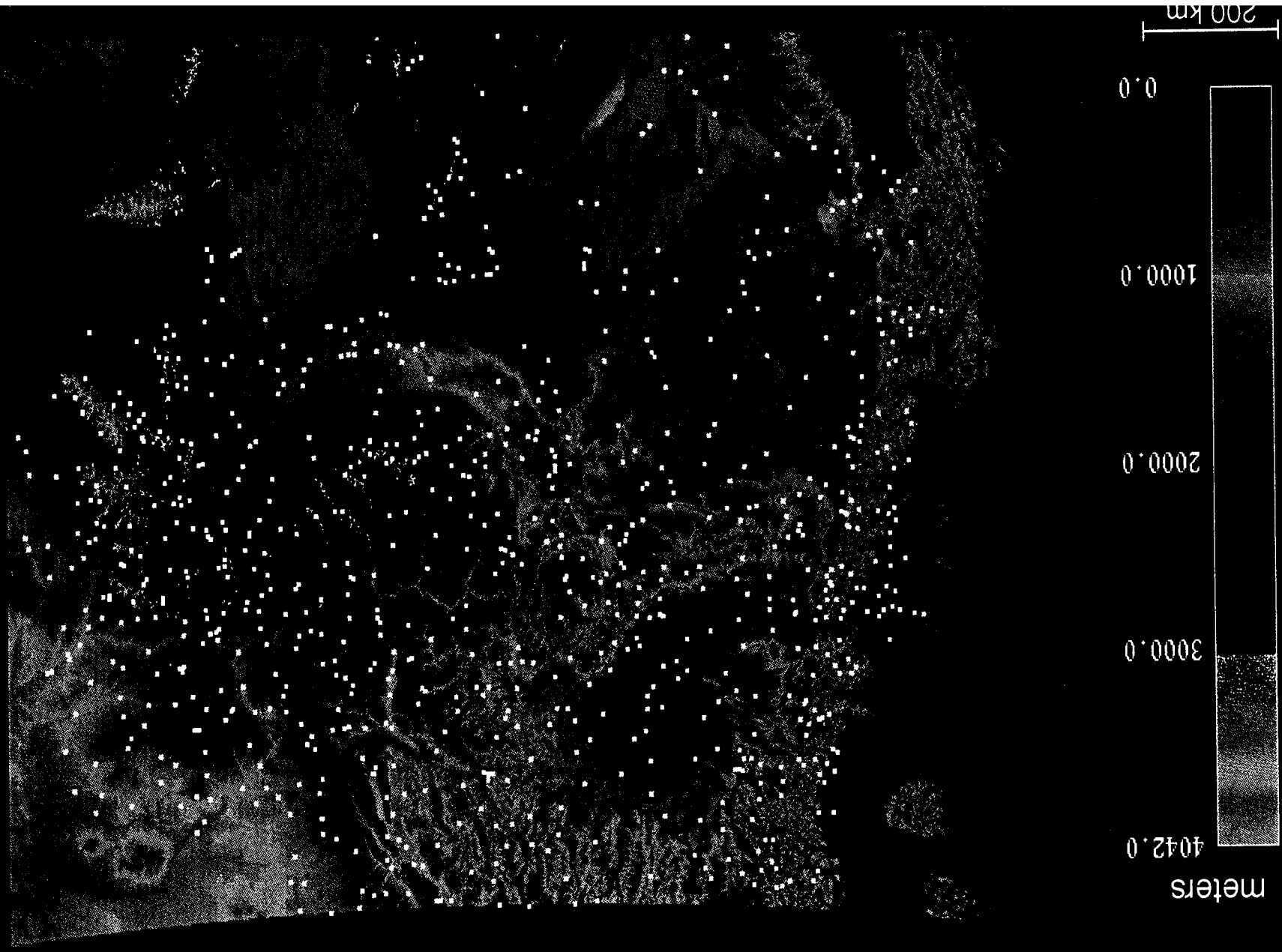
Figure 5. Annual total precipitation over the Columbia Basin study region.

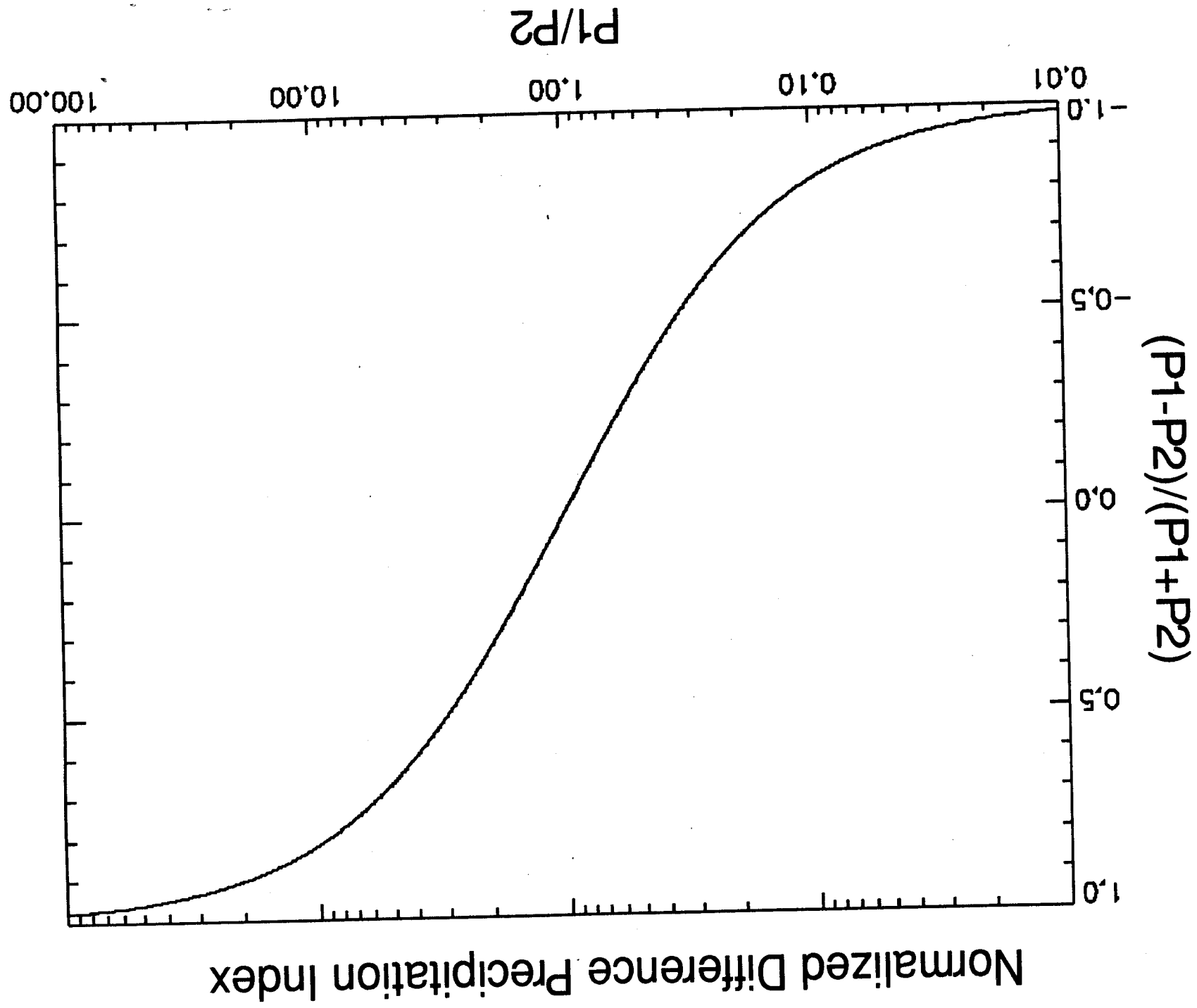
Figure 6. Observed vs. predicted annual average (a) Tmax and (b) Tmin over 456 stations inside the Columbia Basin study region. See Table 1 for regression statistics.

Figure 7. Observed vs. predicted $\ln(\text{annual total Prcp})$ for 62

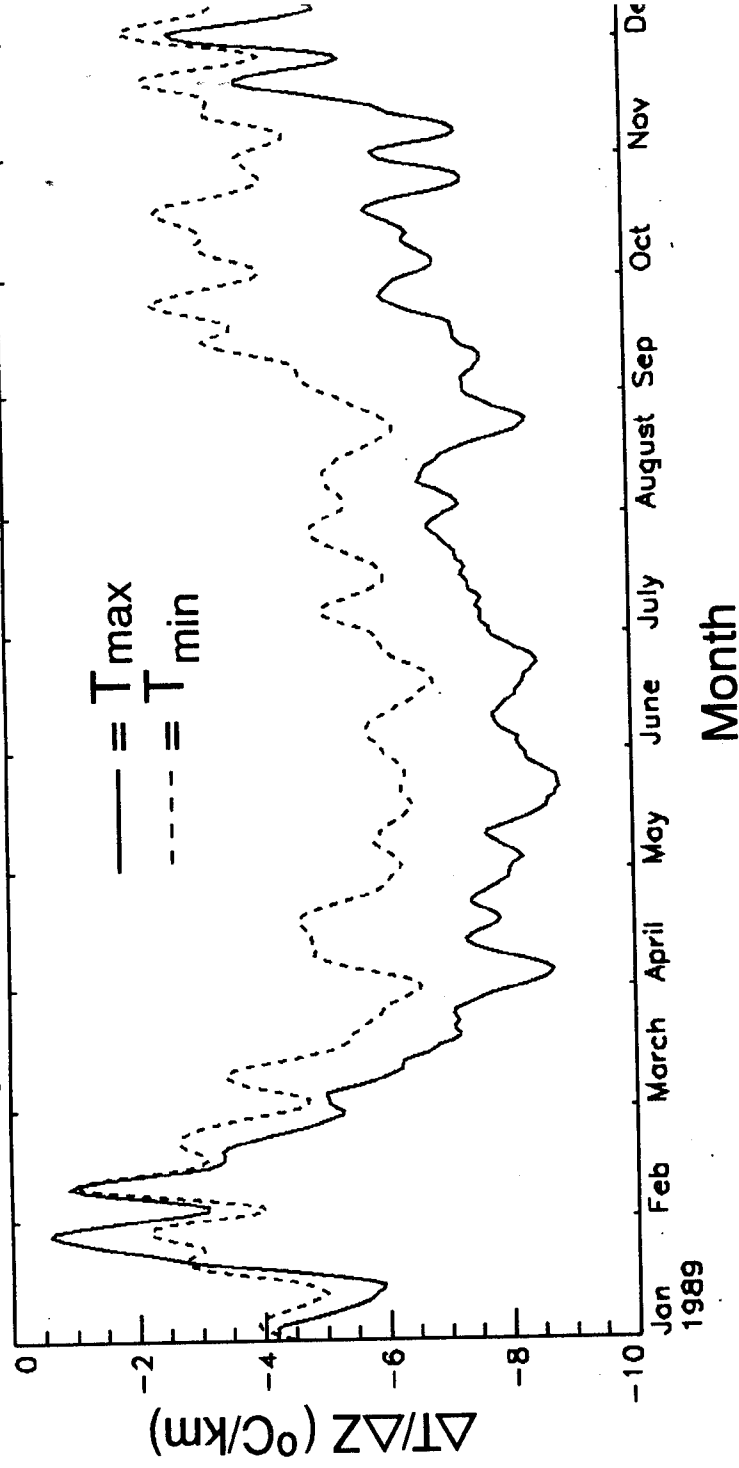
stations inside the Columbia Basin study region. See Table 1 for regression statistics.

Pacific Northwest United States Digital Elevation and Station Markers

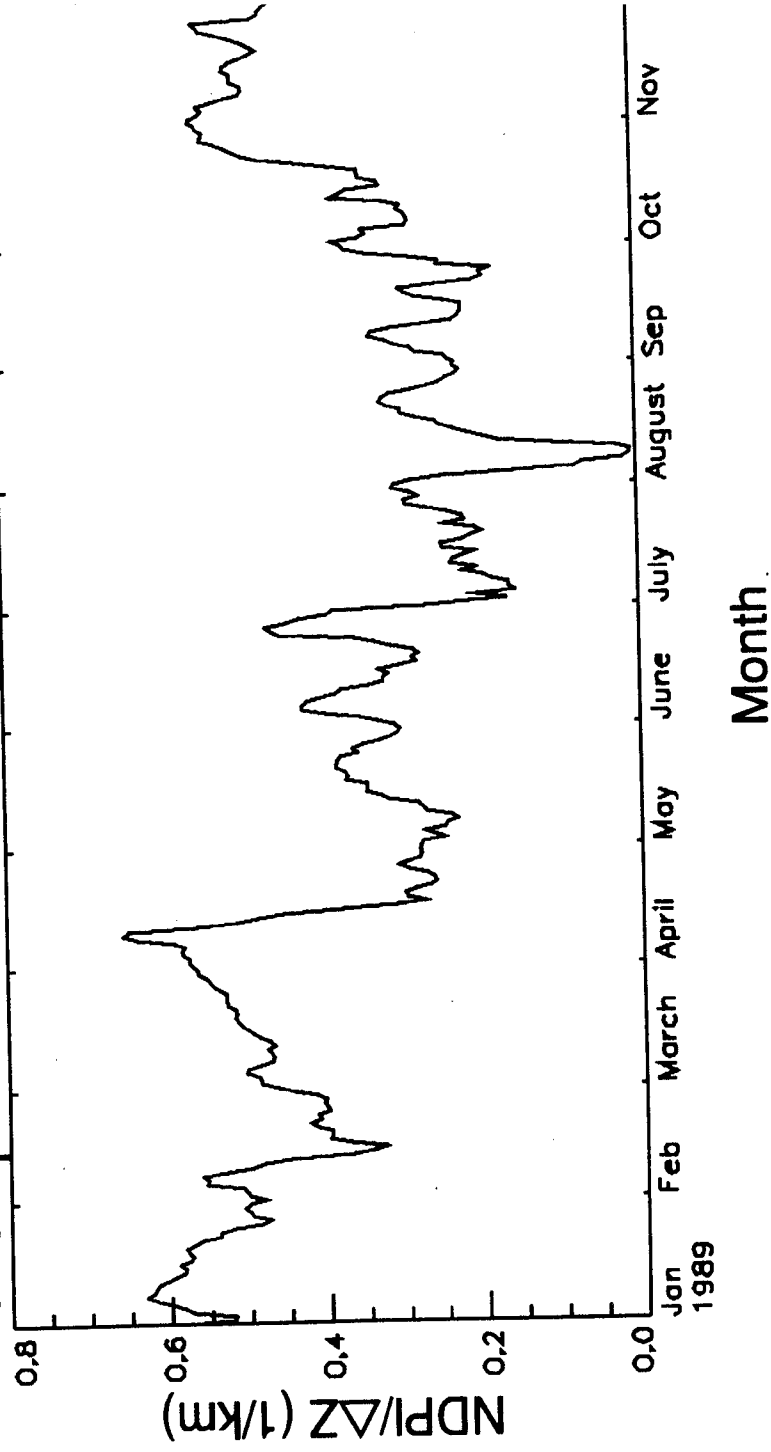




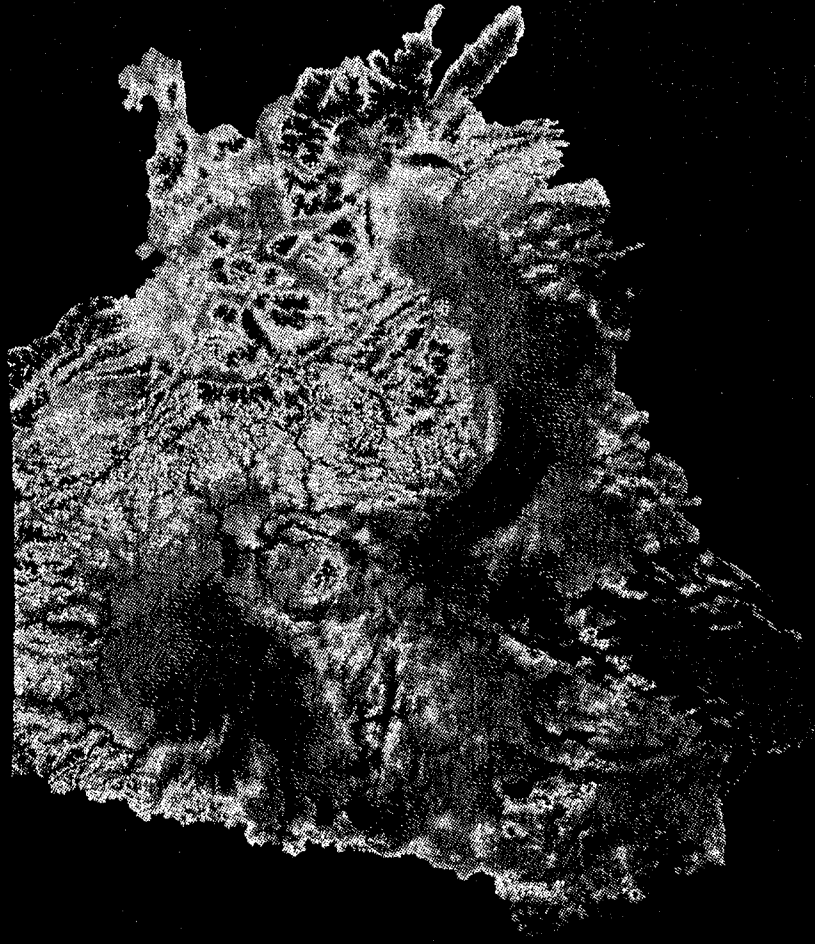
Air Temperature



Precipitation



Annual Average $T_{\max}$, 1989



°C

25.0

15.0

5.0

0.0

-5.0

-15.0

-25.0

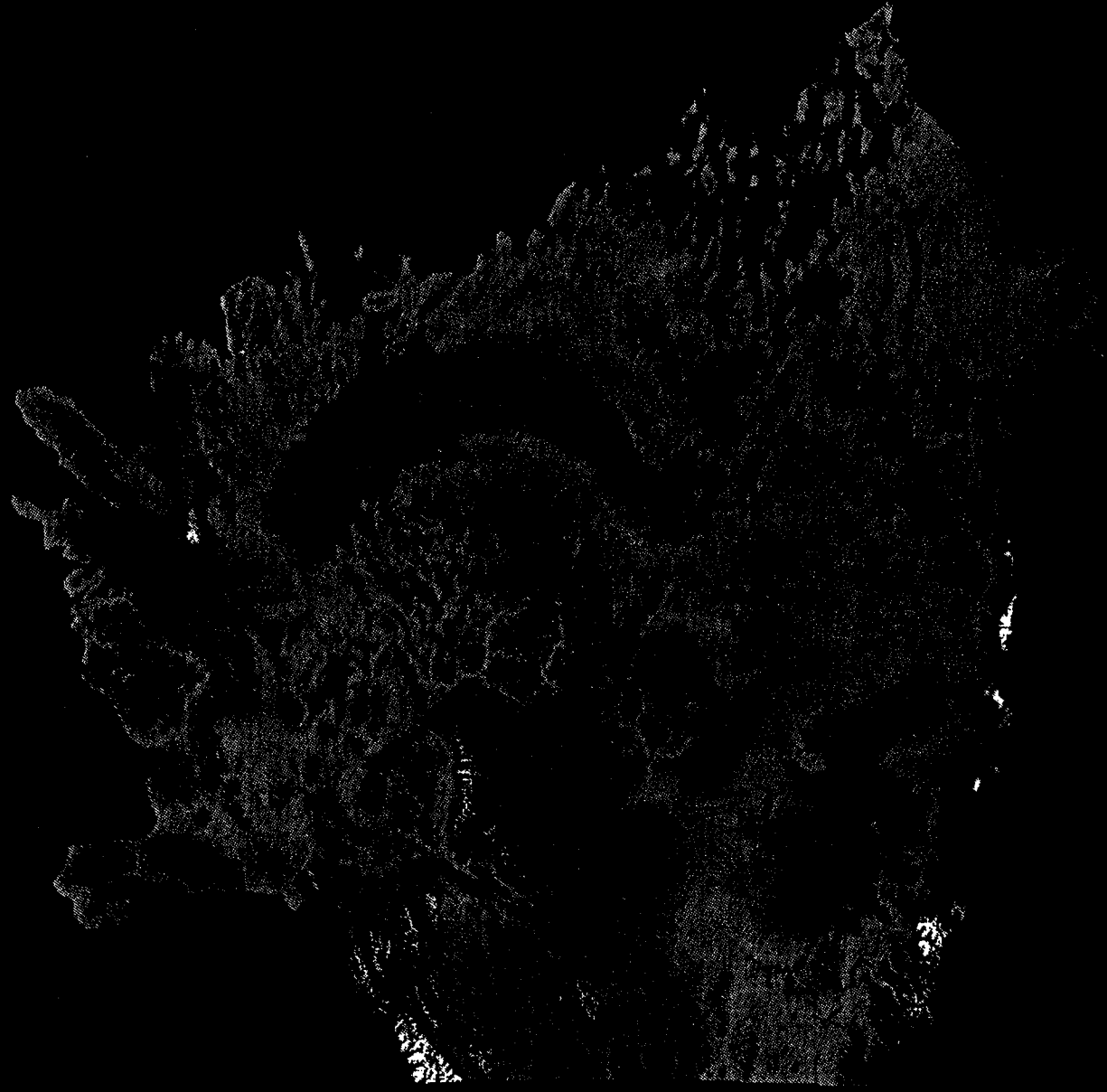


Annual Average $T_{\min}$, 1989



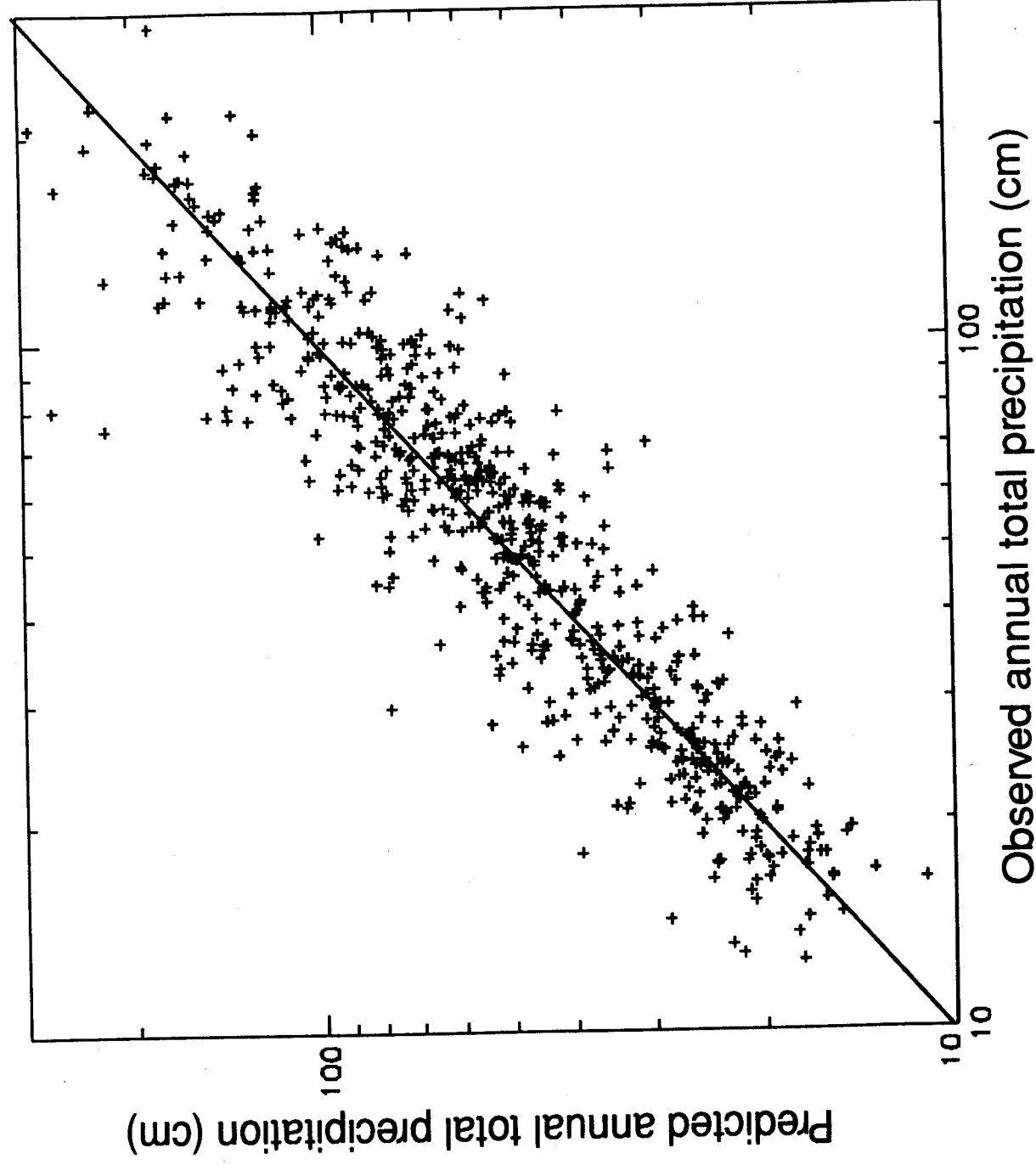
200 km

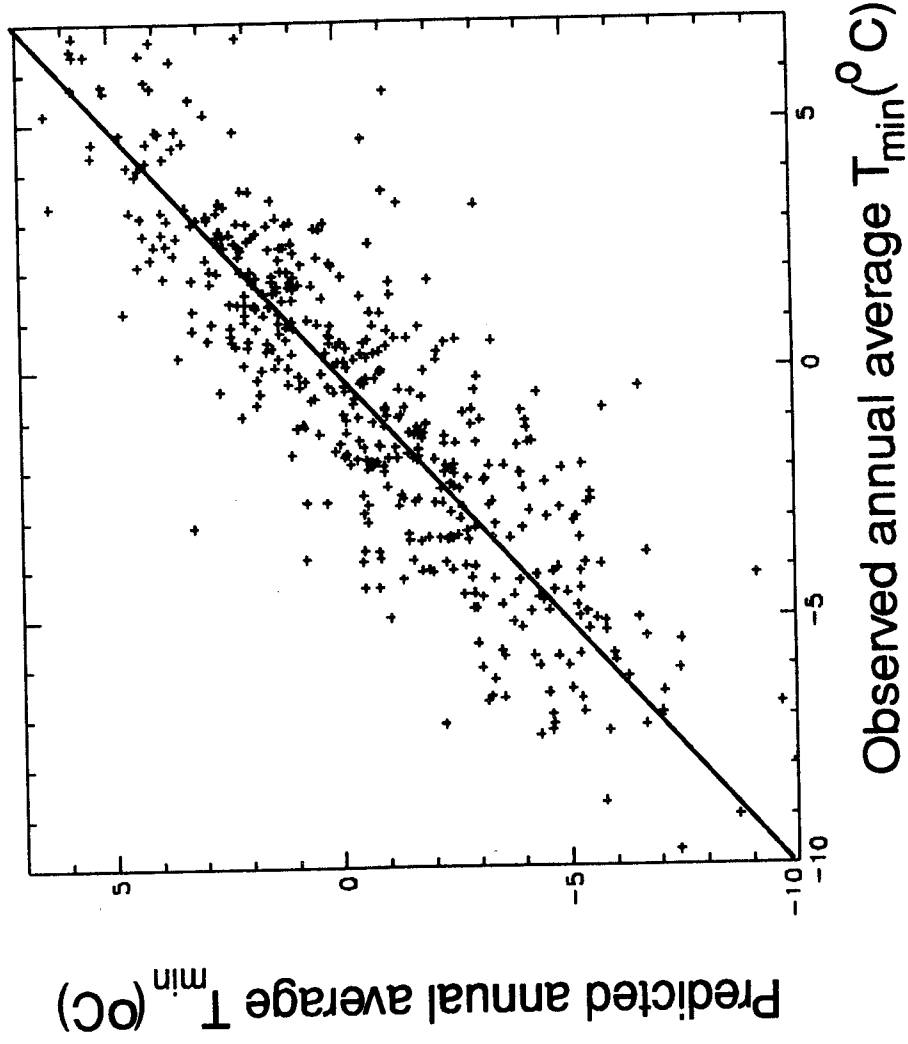
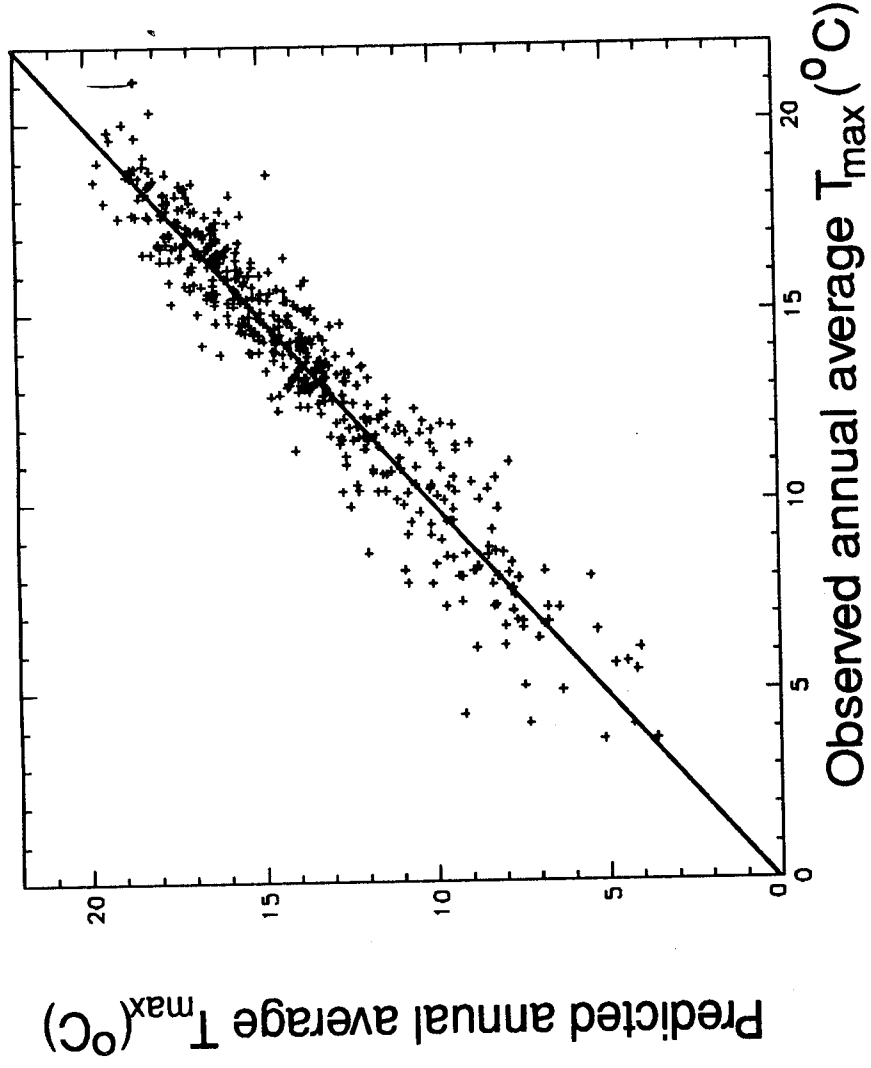
Annual Total Precipitation, 1989



cm
300.0
250.0
200.0
150.0
100.0
60.0
30.0
10.0

200 km





Effects of Soil Moisture Aggregation on Surface Evaporative Fluxes

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Abstract.

The effects of small-scale heterogeneity in land surface characteristics on the large-scale fluxes of water and energy in the land-atmosphere system has become a central focus of many climatology research experiments. The acquisition of high resolution land surface data through remote sensing and intensive land-climatology field experiments (like HAPEX, FIFE, and BOREAS) has provided data to investigate the interactions between microscale land-atmosphere interactions and macroscale models. To determine the effect of small scale heterogeneities, the spatially averaged evaporative fraction is analytically derived for spatially variable soil moisture and soil-atmospheric controls on evaporation at low soil moisture. This average evaporative fraction is compared to the evaporative fraction determined using the spatially averaged soil moisture, as if from a lumped, or aggregated, land surface model. Results show that the lumped-model based evaporation will over estimate evaporation during periods of low atmospheric demands (early morning/late afternoon, Winter periods, etc.) and under estimate evaporation during periods of high demand (mid-day Summer periods.) The accuracy of using 'effective' parameters in lumped macroscale models depends on the variability of soil moisture and the sensitivity of the soil-vegetation system to low soil moisture.

1 Introduction.

The issue of 'scale interaction' for land surface-atmospheric processes has emerged as one of the critical unresolved problems for the parameterization of climate models. To help resolve this issue, the understanding of the scaling properties of water and energy fluxes with their corresponding storage terms (especially soil moisture) has been an important scientific objective of land-climatology experiments like the First ISLSCP Field Experiment (FIFE) (see Sellers et al, 1988) and for new climate projects such as the Global Energy and Water Experiment (GEWEX) Continental project GCIP (see WCRP, 1992). The development of new high resolution 4DDA weather prediction models at NMC (referred to as the Eta model; see Brown, 1994) and a similar model at ECMWF (e.g. see Beljaars and Betts, 1992 or Beljaars and Viterbo, 1994) eventually will incorporate complex surface-vegetation-atmospheric-transfer (SVAT) schemes. Since these models will operate at a 25 to 40 km resolution, the effect of small scale heterogeneity on grid-scale fluxes must be resolved, and if the effects important, how to represent sub-grid variability in such models.

Following Dubayah et al. (1996), for the purposes of this paper, we use the terms 'large scale' or 'macroscale' modeling to correspond to large areas and usually coarse resolution or grid spacing relative to the length scale of the variation in the inputs, processes or model parameters; 'small scale' or 'microscale' modeling to correspond to small areas and usually fine resolution or grid spacing relative to the length scale of the variation in the inputs, processes or model parameters. In this manner, we will refer to the SVAT schemes in a 4DDA model, which assumes homogeneous parameters within the (25 to 40 km) model grid as a 'macroscale' land-atmosphere model.

The acquisition of high resolution land surface data through remote sensing and intensive land-climatology field experiments (like HAPEX and FIFE) has provided initial data sets to investigate the interactions between small scale land-atmosphere interactions and macroscale models. One essential research question is how to account for the small scale heterogeneities, and whether 'effective' parameters and averaged inputs can be used in macroscale models.

The current scientific thinking on this issue is mixed. For example Sellers et al. (1992) state that land-atmospheric models are almost scale invariant, based on analyses of data collected during the First ISLSCP Field Experiment (FIFE); a conclusion also reached by Noilhan et. al. (this issue) using HAPEX-MOBILHY data and suggested by Wood and Lakshmi (1993). Counter arguments have been made by Avissar and Pielke (1989, and subsequent work) who found that heterogeneity in land characteristics resulted in sea-breeze like circulations, and significant differences in surface temperatures and energy fluxes across the patches.

The results of analyses presented later in this paper suggest that soil moisture heterogeneity is the critical variable that controls the non-linear behavior of land-atmospheric interactions, and the effect is most pronounced when soil moisture heterogeneity is such that part of the spatial domain is under soil-vegetation control and part is under atmospheric control. Following Eagleson (1978), by soil-vegetation control, we mean that the atmospheric demand for moisture from the land surface can not be met by evaporation from the soil and transpiration from vegetation. By atmospheric control we mean that the atmospheric demand for moisture from the land surface is less than the maximum rate (under local conditions) that can be delivered by the soil-vegetation system.

From a modeling perspective, it is important to establish the relationship between spatial variability in the inputs and model parameters, the scale being modeled and the proper representation of the hydrologic processes at that scale. The terrestrial water balance, which for a control volume may be written as:

$$\left\langle \frac{\partial S}{\partial t} \right\rangle = \langle P \rangle - \langle E \rangle - \langle Q \rangle \quad (1)$$

where S represents the moisture in the soil column, E evaporation from the land surface into the atmosphere, P the precipitation from the atmosphere to the land surface, and Q the net runoff from the control volume. The spatial average for the control volume is noted by $\langle \rangle$. Equation (1) is valid over all scales and only through the parameterization of individual terms does the water balance equation become a 'distributed' or 'aggregated' model.

A fully 'distributed' model accounts for spatial variability in inputs, processes or pa-

rameters. In most cases some terms in (1) will be aggregated while others will be represented in a distributed manner. The distributed representation can be either deterministic, in which the actual pattern of variability is represented – examples include the European Hydrological System model (SHE) (Abbott et al., 1986a,b) and the 3-D finite element models of Binley et al. (1989) or Paniconi and Wood (1992); or statistical, in which the patterns of variability are represented statistically – examples being models like TOP-MODEL (Beven and Kirkby, 1979) and its variants (see Wood et al, 1990; Famiglietti et al., 1992; Famiglietti and Wood, 1994a; and Wood et al. 1992) in which topography and soil play an important role in the distribution of water within the catchment.

By 'aggregated' or 'lumped-parameter' model we mean a model that represents the catchment (or control volume) as being spatially homogeneous with regard to inputs and parameters. There are a wide number of hydrologic models of varying complexity that do not consider spatial variability. For the runoff term in (1), this includes the well-known SCS Curve Number model and its variants. The water balance models of Eagleson (1978) and the complex GCM atmospheric-biospheric land surface parameterizations like the Biosphere Atmosphere Transfer Scheme (BATS) of Dickinson (1984) and the Simple Biosphere Model (SiB) of Sellers et al. (1986) are aggregated models.

2 Estimation of Evaporation at Large Scales.

2.1 Overview.

The terrestrial water balance, including input forcings, infiltration, evaporation and runoff processes, has been revealed to be a highly nonlinear and spatially variable process. Yet, little progress has been made in relating the observed small-scale complexity that is apparent from recent field and remote sensing experiments to models and parameterizations at large scales. The reason for this lack of progress is due in part to designing these experiments without regard to resolving this scaling question. This manifests itself in the experiments through their limited duration and spatial scales. In addition, the difficulty and cost of collecting the required spatial data, and the lack of process-based distributed

models has further contributed to the limited progress.

Insights into the scaling question, and the potential of averaged or 'equivalent' parameters in lumped, macroscale models is the subject of this paper. In this paper, we will use averaged and equivalent parameters interchangeably, but recognize that the preferred definition of an equivalent parameter is one that results in matching the outputs of a lumped model and a distributed model for example the flux-matching work described in Raupach (1993) and Raupach and Finnigan (1995) Such an equivalent parameter may not be (an often is not) the average of the distributed parameter. Important analyses and insights can be made through detailed model simulation, of the type carried out by Famiglietti and Wood (1995) for the Kings Creek area of FIFE. One limitation of such modeling is its short duration (often one summer season or less) which may not provide the range of conditions necessary to test aggregation hypotheses. Thus, given the nature of the scientific debate concerning aggregation, it is important to develop a more general analysis. In particular, this paper tries to lay out a consistent framework for determining when land surface conditions permit aggregation of subgrid variability and the specification of equivalent parameters, and conditions when such aggregation will fail.

Results presented in section 2.2 show that during periods of low atmospheric demands (early morning/late afternoon, Winter periods, etc.) the use of equivalent parameters will over estimate evaporation; and during periods of high demand (mid-day Summer periods) under estimate evaporation. The accuracy of using equivalent parameters in lumped macroscale models appears to depend on the variability of soil moisture and the sensitivity of the soil-vegetation system to low soil moisture.

2.2 Derived Distribution of Soil Moisture and Evaporation.

In natural watersheds, the spatial heterogeneity in soil, vegetation and topographic properties along with spatial variability in precipitation and net radiation, leads to spatial variability in water table depths, runoff, soil moisture, and evapotranspiration. Understanding the linkage between water table depths, soil moisture and evapotranspiration is necessary for understanding heterogeneity within natural watersheds. Field data shows

that areas with water table depths near the ground surface tend to higher surface soil moisture, in part due to capillary rise and diffusion from the saturated to the unsaturated soil zone. Higher near surface soil moisture increases the maximum rate at which the soil-vegetation system can deliver water to the atmosphere. Eagleson (1978) defined this rate as the 'soil controlled' evapotranspiration rate. For areas with deep water tables, and subsequently dry surface soil moisture conditions, the maximum rate that the soil-vegetation system can deliver is less than the evaporation determined solely from atmospheric conditions. The rate determined from atmospheric conditions is often referred to as the 'atmospheric' demand for water (Eagleson, 1978) and is the potential evapotranspiration rate. Actual evapotranspiration from a natural surface will be the minimum of these two rates: 'soil controlled' and 'atmospheric controlled'. Within a heterogeneous watershed, with a distribution of water table depths, leading to a distribution of surface soil moisture levels, there will be areas of high surface moisture evapotranspiring at the atmospheric controlled rate and areas of low soil moisture evapotranspiring at the soil controlled rate.

The concept of a distributed land surface hydrologic model is to try to account for and represent this variability. Models that represent this variability statistically are referred to in this paper as macroscale, distributed models. For TOPLATS (see Famiglietti and Wood, 1994a) the distribution in the soil-topographic index leads to a distribution in water table depths (see also Beven and Kirkby, 1979; Wood et al., 1990; or Famiglietti et al., 1992). From the distribution of water table depths z , the statistical distribution for surface soil moisture can be derived using soil characteristic relations which relates the soil matrix head, ψ , to soil moisture as a function of soil properties. (Certain assumptions are usually applied, such as gravity drainage $\partial\psi/\partial z = 1$, steady state vertical flow, and uniform soil moisture profiles.) Thus, given the atmospheric evaporative demand and the statistical distribution of soil moisture, the distribution of the actual evapotranspiration, E_a , can be derived. For simple functional forms the mapping of $z \rightarrow \theta \rightarrow E_a$ can be done analytically; and in any case it can be done through simulation.

Figure 1, from Wood (1994), provides some TOPLATS model simulation results to

illustrate how the statistical distribution of the actual soil evaporation can be estimated and by analogy evapotranspiration. Results are provided for two times during the day: an early morning/late afternoon time where the atmospheric evaporative demand (potential rate) is approximately 0.35 mm/hr and a mid-day time with an atmospheric evaporative demand of approximately 0.78 mm/hr. For surface soil moisture levels too low to meet this demand, the actual evaporation function decreases from the potential rate, as shown in panels (iii) of Figures 1a and 1b. The parameters for the curves are taken from Famiglietti and Wood (1994b) and represent conditions for the 12.7 sq. km. Kings Creek watershed within the FIFE experiment area near Manhattan, Kansas.

To derive the probability distribution for the actual evaporation, the probability distribution for water table depth is combined with two relationships: one that relates water table depth to surface soil moisture and a second that relates surface soil moisture to actual evaporation. The first relationship is assumed to be fixed in time. The second relationship changes throughout the day and is constrained at any time (at a maximum potential rate) due to the diurnal cycle in solar radiation. Thus, the probability distribution of water table depths is mapped into a probability distribution for actual evaporation.

Two conditions are simulated: Figure 1a represents quite dry conditions - low water table, and Figure 1b (for the same diurnal times) for a wetter (but not extremely wet) condition. The figures have been divided into four panels that show the derivation of the bare soil evaporation distribution. Panel (i) in the lower right corner gives the probability distribution for water table depths derived from the soil-topographic index of TOPLATS and relies on the basic underlying hydrologic similarity theory of TOPMODEL (Beven and Kirkby, 1979.) Panel (ii) relates water table depth to surface soil moisture as discussed earlier in this section. These two figures could generate a derived probability distribution for surface soil moisture. This is not shown here. The upper left portion of the figure, panel (iii), gives the relationship between surface soil moisture and actual evaporation for the two times during the day. The maximum evaporation rate is the potential rate, which is lower during the early morning and late afternoon (due to lower net radiation.) For areas within the watershed (or grid) where surface soil moisture is high, the actual evaporation

rate is equal to the potential rate. For drier areas, the rate is lower. The resulting probability distribution for the actual evaporation is shown in the lower left portion of the figure. This panel is divided into two; the top, panel (iv-a), giving the distribution for the time related to early morning/late afternoon (lower potential evaporation rate) while the bottom, panel (iv-b), being related to mid-day high potential evaporation condition.

Inspection of the derived probability distributions for the two times (early morning/late afternoon and mid-day) and two moisture conditions reveals that for the very dry conditions the probability distribution for evaporation has a narrow range and the average water table depth can be used to estimate the average evaporation rate. This result arises because the soil moisture - evaporation function is essentially linear in the range of soil moistures representing the dry conditions. For the wetter condition, the function is non-linear and the range of soil moistures within the watershed contains areas which are very dry (having low evaporation rates) and wet (having evaporation rates at potential.) If the conditions were even wetter, then the distribution of evaporation rates would be concentrated at the potential evaporation rate. Thus, these results suggest that the non-linear effects of soil moisture on evaporation (and therefore the largest errors when using spatially averaged parameters) occurs in areas with a wide range of soil moisture conditions (due to variation in soils or drying conditions).

To test this sensitivity to dry soil conditions, comparisons were made between the distributed water-energy balance model (TOPLATS) with a model grid of 30 m and using USGS DEM to represent topography and a lumped representation (one-dimensional model with averaged parameters). The simulations were run for 5 days during the October 1987 FIFE intensive field campaign, IFC-4. This period had the driest conditions observed during the 1987 experiment. Figure 2 (from Famiglietti and Wood, 1995) shows the simulations for October 6 - 11, 1987. The models were run at a 0.5 hour time step to capture the diurnal cycle in potential evapotranspiration. Three models are compared: a fully distributed model in which the spatial patterns of the soil-topographic parameters and rainfall are preserved; a macroscale, distributed model in which the spatial variability in the soil-topographic parameters is accounted for statistically; and an aggregated, one-

dimensional model in which parameters and inputs are spatially constant at their average values.

The one-dimensional model predicts well the evapotranspiration during the morning and late afternoon when the atmospheric demand for moisture (potential evaporation) is low, but fails to accurately predict evaporation during mid-day periods when soil (and vegetation) controls limit the actual evapotranspiration. It is during this period that the sensitivity of evaporation to soil moisture is high. By ignoring the spatial variability in soil moisture the aggregated model significantly underestimates the catchment-scale evapotranspiration. During wet periods, the one-dimensional model may work quite well. These results illustrate the complexity of the problem: distributed versus aggregated land surface parameterizations, since the errors introduced due to aggregation of parameters varies with the state of the system.

3 Analytical Estimate of Evaporative Fraction.

In the previous section through a discussion of Figure 1, a qualitative analysis was presented concerning the mechanisms leading to a non-linear behavior of evaporation and the under estimation of the evaporative flux when aggregation is carried out. Close examination of the concepts around Figure 1 leads to the insight that difficulties in aggregation occur under a combination of two conditions: (i) when a significant portion of the area is under 'soil controlled' evaporation rates due to a combination of low mean soil moisture and/or large variations in soil moisture, and (ii) when soil (or vegetation) properties result in a large decrease in evaporation (transpiration) under water stressed situations. In actual circumstances, it is an interplay between the soil moisture variability and the soil-vegetation response at low soil moisture that results in the non-linear behavior, resulting in errors with parameter aggregation.

To further explore the errors in estimated evapotranspiration, we develop in this section a simple analytical model for evaporation, and use the model to investigate the sensitivity of estimated evaporation to soil moisture aggregation. The simple analytical model allows us to investigate a wider range of conditions than would be possible through

complex simulation, yet still preserve the essential features of the aggregation problem. For more complicated, distributed models similar results can be simulated as shown in the previous section or, for example, in Famiglietti and Wood (1995).

3.1 Analytical Model for Soil Variability and Evaporation.

Let us assume that at a limiting soil moisture level θ^* , the atmospheric evaporative demand (i.e. potential evaporation) and the ability of the soil/vegetation system to provide moisture to the atmosphere are in equilibrium; i.e. a land-atmospheric equilibrium with regard to soil moisture. (This limiting soil moisture level, θ^* , varies with net radiation and has a diurnal and seasonal cycle.) The fraction of the watershed (or grid) with soil moisture less than θ^* is defined as F , and will be referred to as the fraction of the watershed under 'soil-controlled' evaporation since the evaporation from these areas is limited by the soil-vegetation system. The fraction of the watershed $1 - F$, is under 'atmospheric-controlled' evaporation and the evaporation rate is at the potential rate.

Let the ratio of actual evaporation to potential evaporation for soil moisture, θ , less than θ^* be represented as

$$\begin{aligned} \frac{E_a}{E_p} &= \exp \{-\beta(\theta^* - \theta)\}, & \theta &\leq \theta^* \\ \frac{E_a}{E_p} &= 1, & \theta &\geq \theta^* \end{aligned} \quad (2)$$

For this analysis, β will not depend on θ^* . The values of β will depend on the soil-vegetation characteristics and can be estimated through field work; for example lysimeters. Soil moisture is assumed to vary spatially, and we will approximate its probability distribution (at any time), with a gamma distribution of the form

$$f(\theta) = \frac{\lambda^k (\xi - \theta)^{k-1}}{\Gamma(k)} \exp \{-\lambda(\xi - \theta)\} \quad (3)$$

where ξ is the upper limit on soil moisture and can be taken to be the porosity. (This distribution does not consider time explicitly but can be easily incorporated by making the parameters of (3) time varying.)

3.2 Derivation of the Evaporative Fraction.

The expected value of E_a is computed for both the distributed soil moisture model and for the aggregated model. The comparisons of their expected values will be used to evaluate the accuracy of the aggregated modeling approach. The expected value of E_a/E_p is by definition:

$$\langle \frac{E_a}{E_p} \rangle = \int_{\theta_r}^{\theta^*} \exp \{ -\beta(\theta^* - \theta) \} f(\theta) d\theta + (1 - F_s) \quad (4)$$

where θ_r is the residual soil moisture, and F_s is the fraction of the area that is under 'soil control' and $(1 - F_s)$ the fraction under 'atmospheric control'. Assuming that the cumulative probability for $\theta < \theta_r$ is small, equation (4) can be solved and is equal to

$$\langle \frac{E_a}{E_p} \rangle = (1 - F_s) + \left(\frac{\lambda}{\lambda + \beta} \right)^k \exp \{ -\beta(\theta^* - \theta) \} G(\theta^*, k, \lambda + \beta) \quad (5)$$

where

$$G(\theta^*, k, \lambda + \beta) = \int_{-\infty}^{\theta^*} \frac{(\lambda + \beta)^k (\xi - \theta)^{k-1}}{\Gamma(k)} \exp \{ -(\lambda + \beta)(\xi - \theta) \} d\theta \quad (6)$$

The actual evaporation for the macroscale model with aggregated parameters, $\langle \frac{\bar{E}_a}{E_p} \rangle$, is estimated using the average soil moisture $\langle \theta \rangle$ in equation (2); i.e.

$$\langle \frac{\bar{E}_a}{E_p} \rangle = \exp \{ -\beta(\theta^* - \langle \theta \rangle) \} \quad (7)$$

Errors in using an aggregated or lumped model, instead of the distributed model with subgrid variability can be expressed as the ratio of equation (7) to equation (5). In particular we refer to this ratio, as the *evaporation scaling ratio*, since it relates the effects of scaling evaporation due to model aggregation to scaling evaporation due to averaging of the spatial process. When this ratio is near unity, the effects of using an aggregated model is minimal. This ratio will be investigated for a range of conditions reflective of field conditions.

3.3 Sensitivity of Land Characteristics on the Evaporation Scaling Ratio.

The sensitivity of the evaporation scaling ratio depends on three parameters: the ratio of the equilibrium soil moisture θ^* to the mean soil moisture $\langle \theta \rangle$, $\theta^*/\langle \theta \rangle$; the soil-vegetation

sensitivity factor to soil moisture stress, β , and the variability of soil moisture within the region measured by its coefficient of variation, Cv .

As discussed earlier, θ^* represents the soil moisture level at which the soil-vegetation system is in equilibrium with the atmospheric evaporative demand. Thus, areas with soil moisture lower than θ^* will exhibit a soil-vegetation control and have an actual evaporative rate lower than the potential rate. During early morning hours, θ^* will be low because the atmospheric demand is low; during the day θ^* will increase; this is shown qualitatively in Figure 1 (panel iii).

The parameter β represents the reduction in evaporation with decreases in soil moisture below the equilibrium soil moisture. If the ratio E_a/E_p decreases to 0.5 with a decrease in soil moisture of about 7% below θ^* , then β has a implied value of 0.10. Larger values of β implies a system in which the soil-vegetation is more sensitive to decreases in soil moisture below the equilibrium threshold; for example clay dominated soil systems which have a high resistance to evaporation when dry.

Variation in soil moisture at relatively small scales, on agricultural fields, has been well studied (Bell et. al., 1980; and the review in Beven, 1983). These studies suggest that the variation in soil moisture, as expressed in the coefficient of variation, Cv , is in the range of 0.15 - 0.35. For soil moisture remote sensing using the ESTAR passive microwave airborne instrument, Jackson (personal communication, 1994) reported coefficient of variations at a field scale ranging from 0.29 to 0.40. These results were over the range of soil moisture conditions found during a June 1992 remote sensing experiment in the Little Washita, Oklahoma, basin, a 525 sq. km. catchment in the southern Great Plains region of the USA. At the beginning of the experiment the soil moisture was close to saturation and over the next 10 rain-free days experienced a strong dry down. During a subsequent 1994 Shuttle Imaging Radar soil moisture experiment in the same basin, field (0.25 to 0.5 sq. mi.) observed soil moisture Cv (from approximately 20 to 25 samples per field) ranged from 0.11 (for winter wheat in dry conditions) to 0.44 for natural pasture after a light rain. The five winter wheat fields ranged from 0.11 to 0.24, the five pasture fields from 0.15 to 0.44 and two bare soil fields (plowed) from 0.19 to 0.28. The fields have uniform

vegetation type and soil texture class.

A series of sensitivity experiments were conducted by varying these three parameters: $\theta^*/\langle\theta\rangle$ within the range of 0.2 to 2; β within the range 0.05 to 0.6; and Cv within the range 0.10 to 0.40. The results are provided in Figures 3 and 4. Figure 3 provides the sensitivity to changes in β for different Cv values, while Figure 4 provides the sensitivity to changes in Cv for different levels of β . By inspecting Figures 3 and 4, it appears that the following conclusions can be reached:

(i) For all cases, when the average soil moisture exceeds the equilibrium soil moisture, the lumped, aggregated model over estimates the evaporative flux. This over estimation is a maximum when the equilibrium soil moisture is equal to the spatial mean soil moisture and occurs because the aggregated model fails to consider the drier soil areas that evaporate at lower rates. As β increases, these areas are more sensitive to lower soil moisture; thus, this over estimation increases with soil-vegetation systems with larger β values. The over estimation will occur during early and late periods of the diurnal cycle when the atmospheric demand is low resulting in a low θ^* . It will also occur during late Fall-Winter-early Spring periods of low atmospheric evaporative demand and during wet periods when $\langle\theta\rangle$ is large. The overestimation can range up to 50%.

(ii) For cases where the average soil moisture is less than the equilibrium soil moisture, ($\theta^*/\langle\theta\rangle > 1$), the evaporative scaling coefficient is less than 1.0 implying that the lumped, aggregated model underestimates the areal averaged evaporation. This occurs for similar, but opposite, reasons as for the over estimation when $\langle\theta\rangle$ was large relative to θ^* ; namely, the aggregated model fails to consider wetter areas that are evaporating at higher rates. Again, the biggest sensitivity is for soil-vegetation systems with large values of β , and for conditions of large variations in soil moisture. The under estimation will be greatest during mid-day Summer periods when atmospheric evaporative demands are high and surface soil moisture low. Results indicate that the aggregated model underestimates evaporation in the range of 15% to 50% relative to the modeled evaporation from the distributed model for similar conditions.

(iii) For cases of low β and low variation in soil moisture, the evaporative scaling

coefficient is close to 1.0, showing that the aggregated models and the distributed models estimate the same evaporation.

These results imply that aggregated, lumped parameter models will rarely compute evaporation that is equal to that computed by a distributed model; even under apparently 'atmospheric controlled' conditions. This conclusion is due to the variability in soil moisture found within natural landscapes.

4 Discussion and Conclusions

The effect of subgrid variability in soil moisture on evaporation has been investigated with the aim of resolving whether effective (or average) values for soil moisture can replace the distribution found within a catchment or GCM grid square. The simulation results from Kings Creek watershed for the FIFE'87 period (Famiglietti and Wood, 1995) discussed in section 2.2 and the analytical results presented in section 3 suggest that under many natural conditions significant errors in model-estimated evapotranspiration will arise from the use of spatially-averaged soil moisture. This raises the question of how to approach this problem.

The brute force approach would be to incorporate distributed land surface-atmospheric models into climate models and 4DDA weather prediction models. The computational difficulties in applying such distributed models may be quite reasonable if the distributed models represented the small scale variability statistically; i.e. take a TOPMODEL or VIC-2L approach. Still be be resolved in the 4DDA and GCM models is the issue of sub-grid rainfall variability. This takes on increased importance once it is recognized that sub-grid soil moisture must be accounted for.

Another avenue of research is exploring whether precipitation and soil moisture fields scale statistically; using the concepts of statistical self-similarity (see Gupta and Waymire, 1990, for an overview for rainfall and river flows, and Gupta and Dawdy (1994) for its application to flood peaks.) If soil moisture fields can be shown to scale statistically then its small scale statistical properties can be estimated and combined with large scale *mean* soil moisture estimated from either remote sensing or aggregated parameter macroscale

models.

Dubayah et al., (1996) performed a statistical self-similarity analysis using soil moisture fields estimated from an airborne remote sensing passive microwave (ESTAR) instrument. The data was collected over the Little Washita, OK watershed during June 1992. Results show that the soil moisture fields scale, but in a non-linear manner; i.e. violates simple scaling (Gupta and Waymire, 1990.) Continued research into statistical self-similarity for soil moisture suggests great promise for scaling soil moisture. The approach, if successful, would develop a consistent framework for combining soil moisture scaling theory with large scale estimates from 'effective' parameter macroscale hydrological models. The scaling theory would determine the small scale statistical behavior of soil moisture that would then be used to estimate the surface evaporative fluxes through a statistically distributed land surface model.

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Figure 1b: Schematic of deriving the probability distribution for evaporation from the distribution of water table depths for wet conditions (average water table depth is 2.0 m). (i) distribution of the water table depths, (ii) surface soil moisture as a function of the depth of the water table, (iii) actual bare soil evaporation as a function of surface soil moisture for (a) early morning with low potential evaporation and (b) mid-day with high potential evaporation, (iv) derived actual evaporation distributions for early morning conditions (iv-a) and mid-day (iv-b).

Figure 2: Comparisons between two distributed TOPLATS evapotranspirations simulations and a one-dimensional representation for October 6-11, 1987 of the FIFE'87 experiment (from Famiglietti and Wood, 1993.)

Figure 3: Evaporation scaling coefficient versus the ratio of equilibrium soil moisture to areal averaged soil moisture for β ranging from 0.05 to 0.60. (i) $Cv = 0.10$, (ii) $Cv = 0.15$, (iii) $Cv = 0.20$, (iv) $Cv = 0.40$.

Figure 4: Evaporation scaling coefficient versus the ratio of equilibrium soil moisture to areal averaged soil moisture for Cv ranging from 0.10 to 0.40. (i) $\beta = 0.05$, (ii) $\beta = 0.10$, (iii) $\beta = 0.40$, (iv) $\beta = 0.80$.

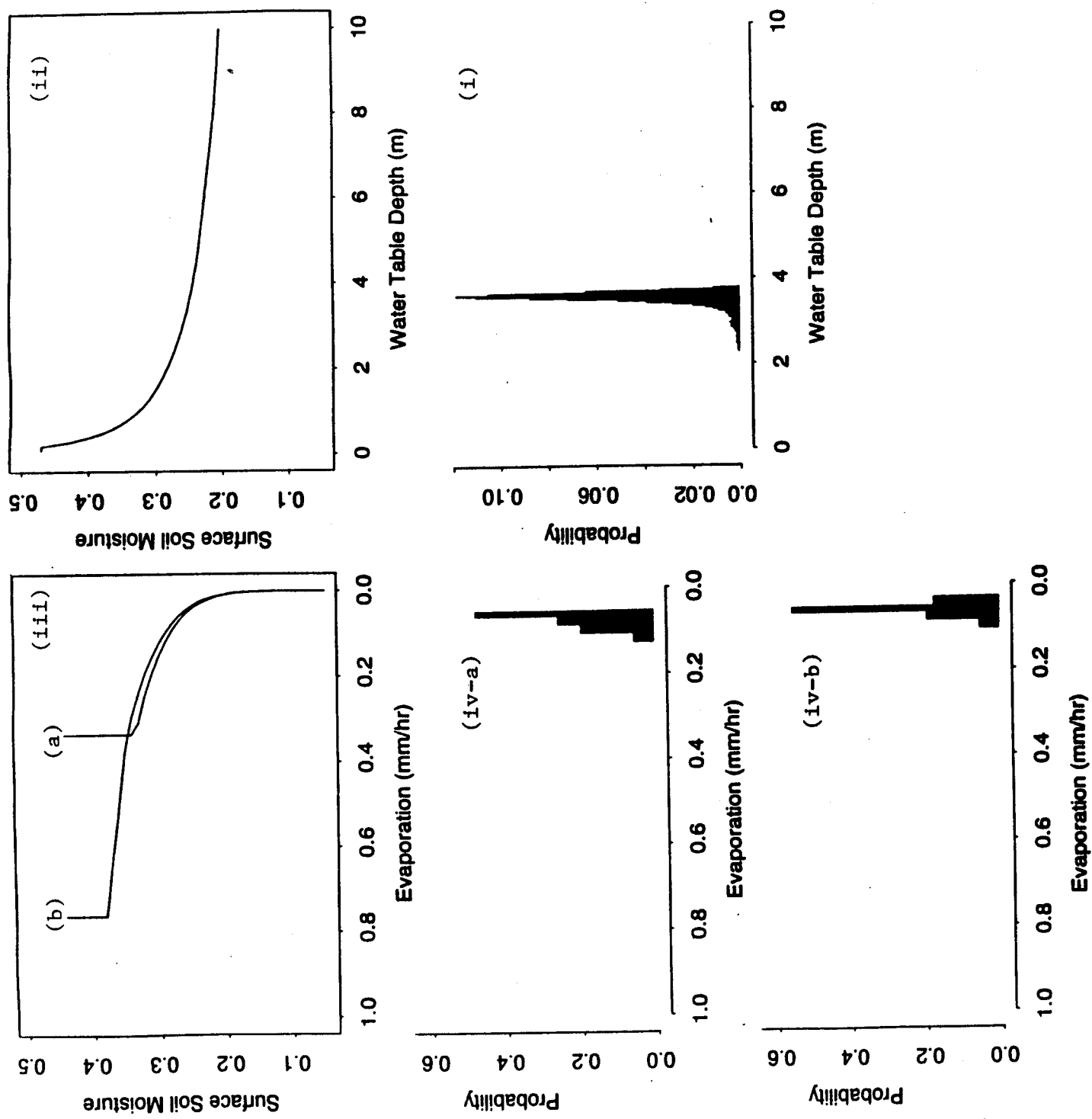


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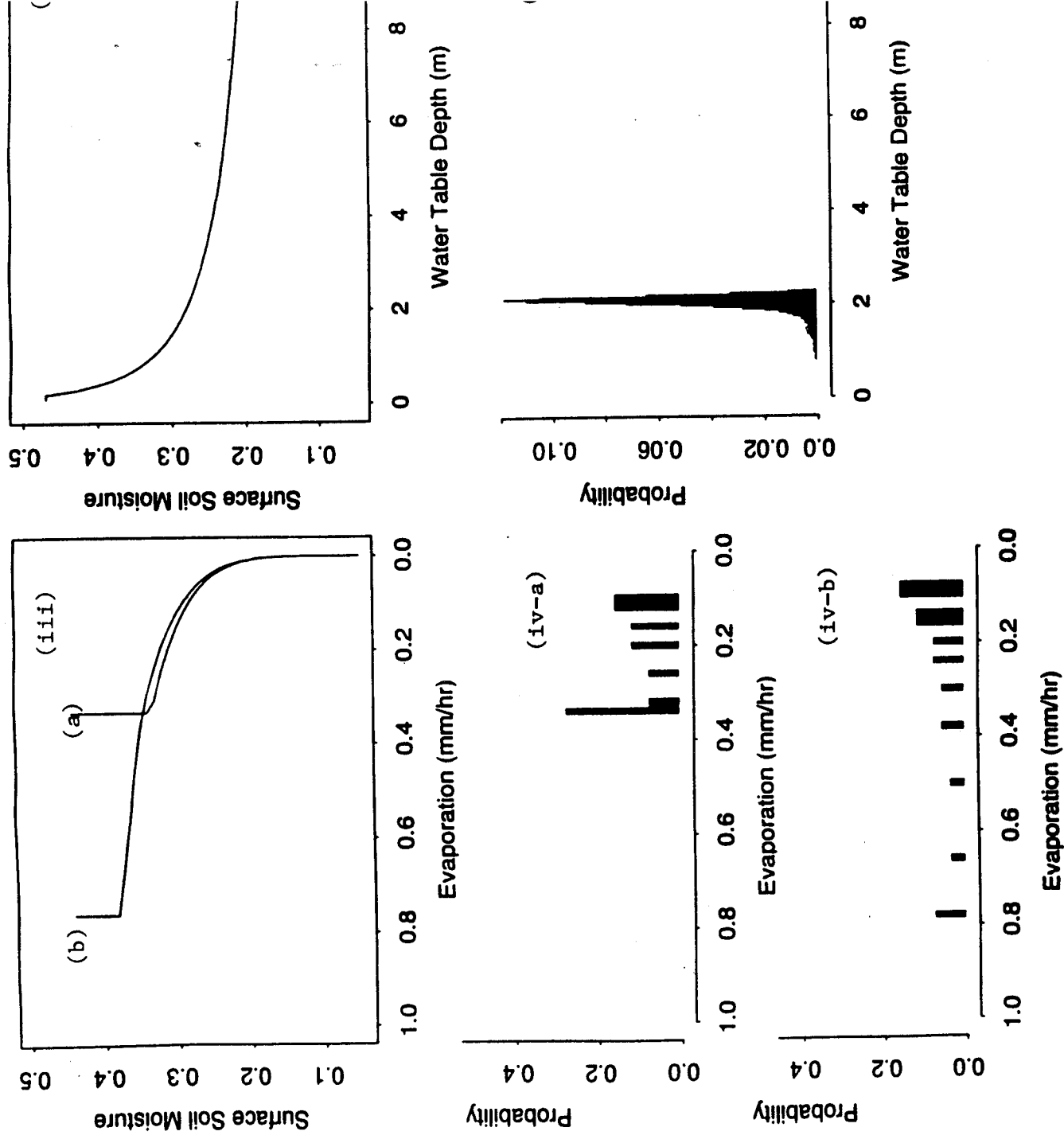


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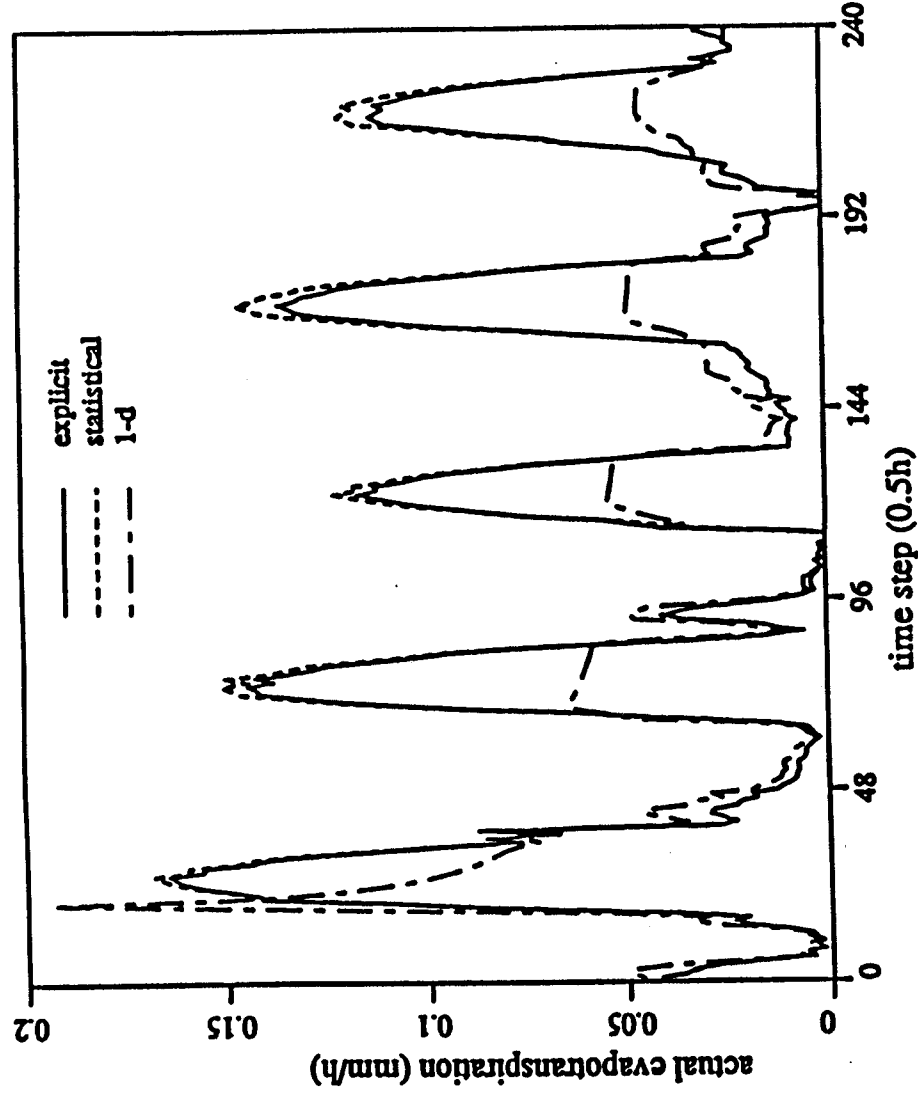


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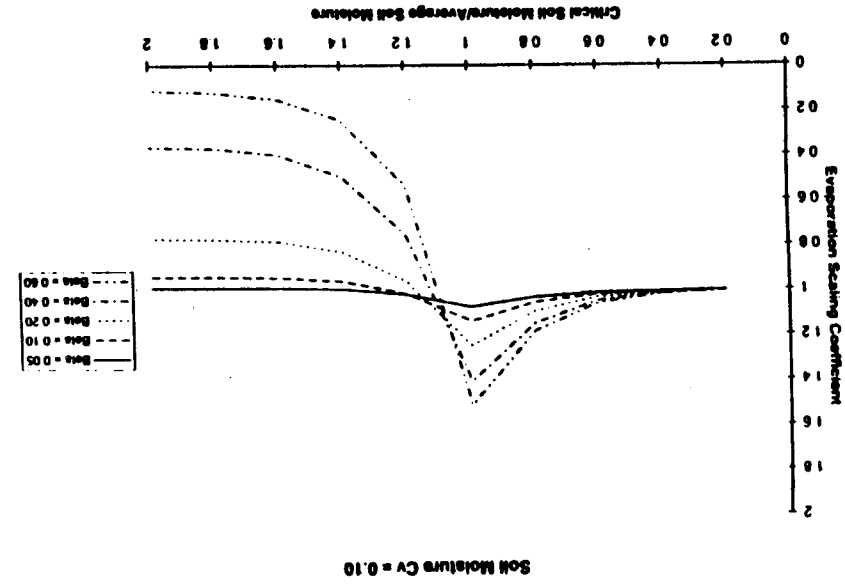
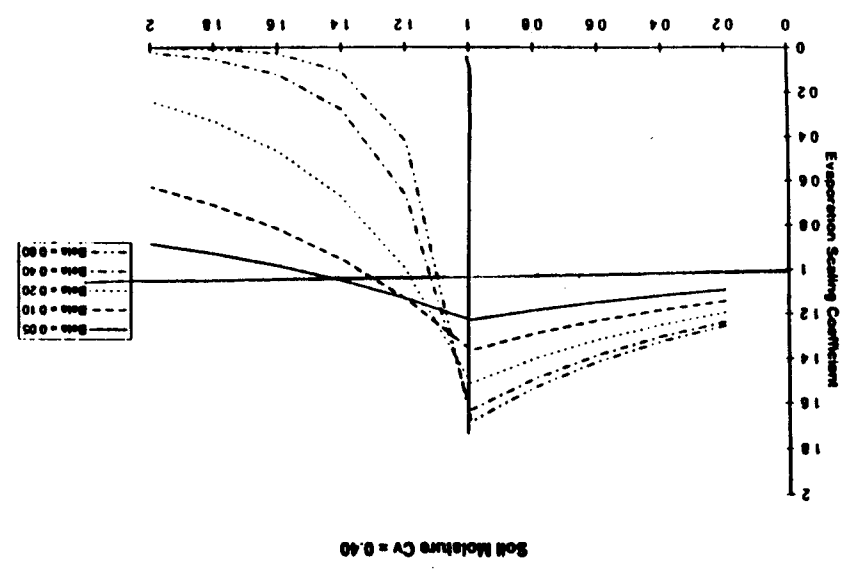
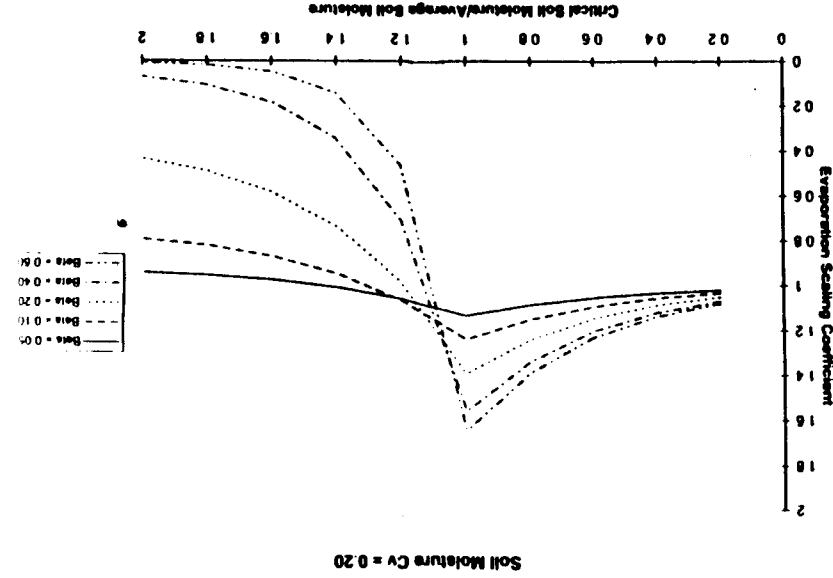
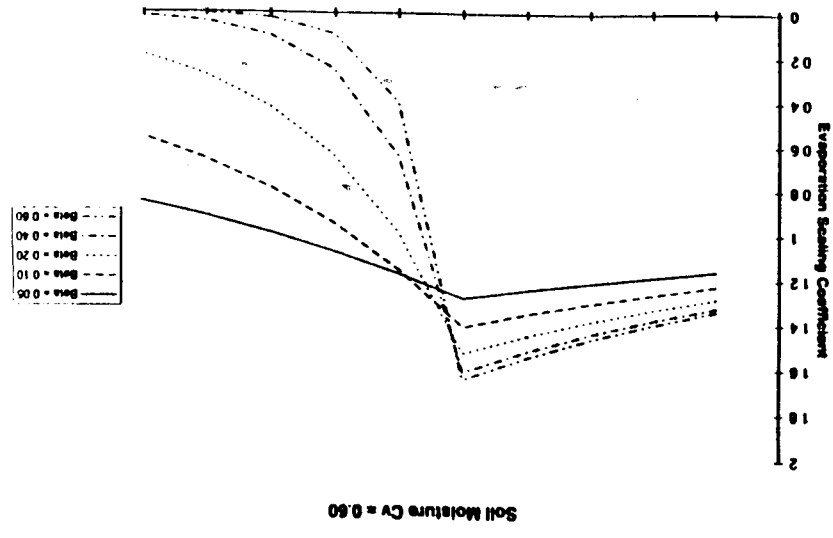


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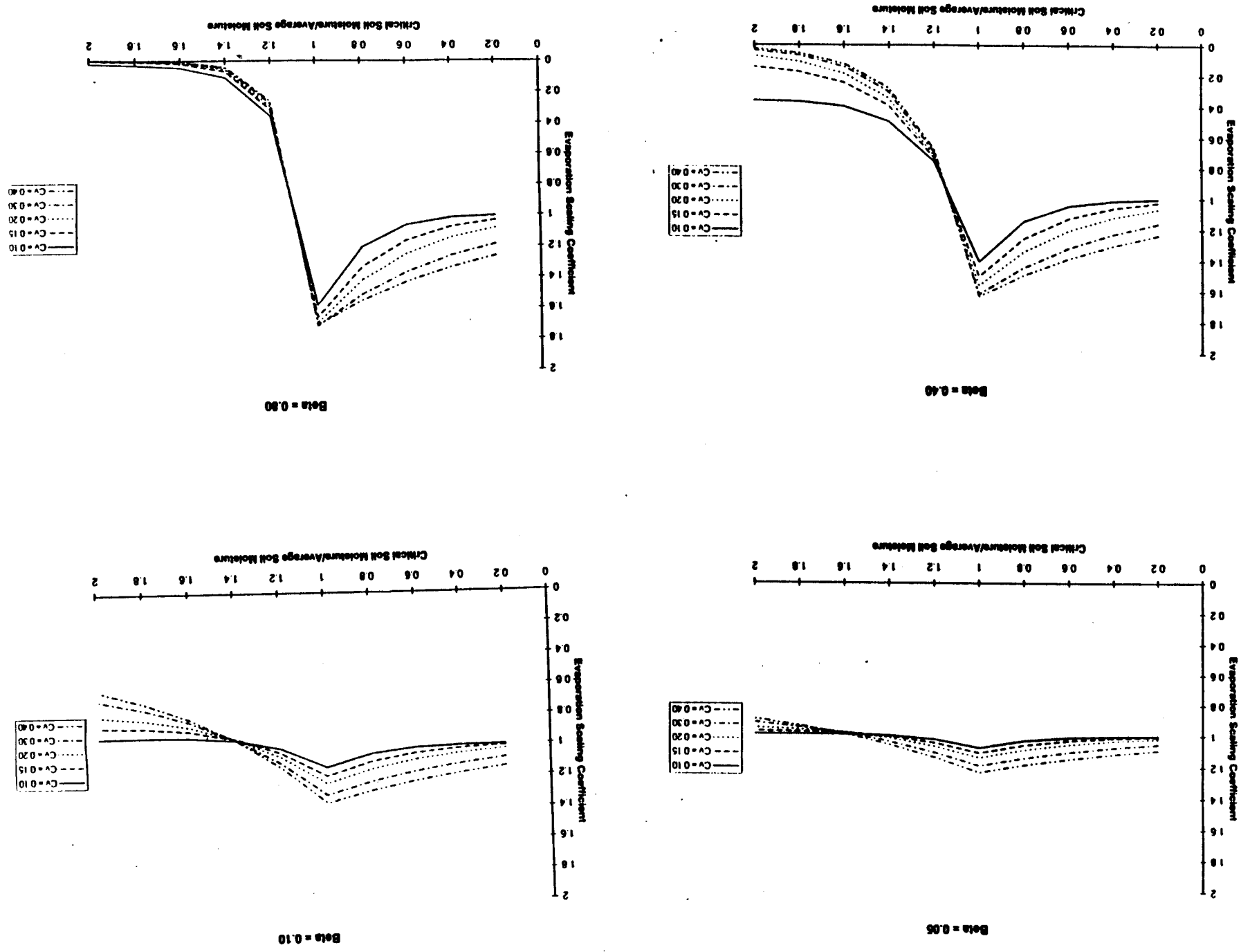


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